TECTONICS, VOL. 32, 171–190, doi:10.1002/tect.20025, 2013

Orogenic plateau growth: Expansion of the Turkish-Iranian Plateau across the Zagros fold-and-thrust belt

M. B. Allen,[1] C. Saville,[1] E. J-P. Blanc,[2] M. Talebian,[3] and E. Nissen[4]

Received 9 July 2012; revised 27 November 2012; accepted 3 February 2013; published 26 April 2013.

[1] This paper shows how the Turkish-Iranian Plateau grows laterally by incrementally incorporating adjacent parts of the Zagros fold-and-thrust belt. The limit of significant, seismogenic, thrusting in the Zagros ($M_w > 5$) occurs close to the regional 1250 m elevation contour. The seismicity cutoff is not a significant bedrock geology boundary. Elevations increase northward, toward regional plateau elevations of ~2 km, implying that another process produced the extra elevation. Between the seismogenic limit of thrusting and the suture, this process is a plausibly ductile thickening of the basement, suggesting depth-dependent strain during compression. Similar depth-dependant crustal strain may explain why the Tibetan plateau has regional elevations ~1500 m greater than the elevation limit of seismogenic thrusting at its margins. We estimate ~68 km shortening across the Zagros Simply Folded Belt in the Fars region, and ~120 km total shortening of the Arabian plate. The Dezful Embayment is a low strain zone in the western Zagros. Deformation is more intense to its northeast, in the Bakhtyari Culmination. The orogenic taper (across strike topographic gradient) across the Dezful Embayment is 0.0004, and across the Bakhtyari Culmination, 0.022. Lateral plateau growth is more pronounced farther east (Fars), where a more uniform structure has a taper of ~0.010 up to elevations of ~1750 m. A >100 km wide region of the Zagros further northeast has a taper of 0.002 and is effectively part of the Turkish-Iranian Plateau. Internal drainage enhances plateau development but is not a pre-requisite. Aspects of the seismicity, structure, and geomorphology of the Zagros do not support critical taper models for fold-and-thrust belts.

Citation: Allen, M. B., C. Saville, E. J.-P. Blanc, M. Talebian, and E. Nissen (2013), Orogenic plateau growth: Expansion of the Turkish-Iranian Plateau across the Zagros fold-and-thrust belt, *Tectonics*, *32*, 171–190, doi:10.1002/tect.20025.

1. Introduction

[2] The purpose of this paper is to examine the boundary and relationships between the Turkish-Iranian Plateau and the Zagros fold-and-thrust belt to its south (Figure 1), thereby improving our understanding of how and why such orogenic plateaux grow in general. This subject is relevant to the wider issue of how and why the continents deform as they do. We show that plateau growth across the Zagros takes place incrementally, that the present limit of seismogenic thrusting does not coincide with a major boundary in the bedrock geology or a break in the regional topographic slope, and that plateau formation takes place after and to the northeast of the cutoff of thrust seismicity—implying that another process thickens the crust at elevations higher than the seismogenic limit. These relationships are relevant to the current debate as to whether continuum or microplate models are most appropriate for continental deformation, including the Arabia-Eurasia collision [*Reilinger et al.*, 2006; *Liu and Bird*, 2008].

[3] Orogenic plateaux develop as a result of continent-continent collision, as demonstrated by the Turkish-Iranian Plateau within the Arabia-Eurasia collision zone [*Şengör and Kidd*, 1979; *Hatzfeld and Molnar*, 2010]. A plateau can form during continental collision where crustal thickening and surface uplift are combined with relatively low erosion and incision rates that limit further thickening and relief generation [*Liu-Zeng et al.*, 2008]. The Altiplano-Puna of South America shows that orogenic plateau formation can also occur without continental collision, during oceanic subduction.

[4] Elevated crust possesses more gravitational potential energy than surrounding lowlands, leading to a buoyancy force that resists further thickening and elevation of the plateau [*England and Houseman*, 1988]. Consistent with this idea, the interior of the Turkish-Iranian Plateau is not undergoing significant upper crustal shortening or thickening at present, indicated by the scarcity of active thrusts [*Talebian and Jackson*, 2004; *Dhont et al.*, 2006; *Allen et al.*, 2011a] and low internal strain, indicated by the Global Positioning System (GPS)-derived velocity field [*Vernant et al.* 2004]. The simple explanation for the relationship between

[1]Department of Earth Sciences, University of Durham, Durham, UK.
[2]Statoil, Oslo, Norway.
[3]Research Institute for Earth Sciences, Geological Survey of Iran, Tehran, Iran.
[4]Department of Geophysics, Colorado School of Mines, Golden, Colorado, USA.

Corresponding author: Mark B. Allen, Department of Earth Sciences, University of Durham, Durham, United Kingdom. (m.b.allen@durham.ac.uk)

0278-7407/13/10.1002/tect.20025

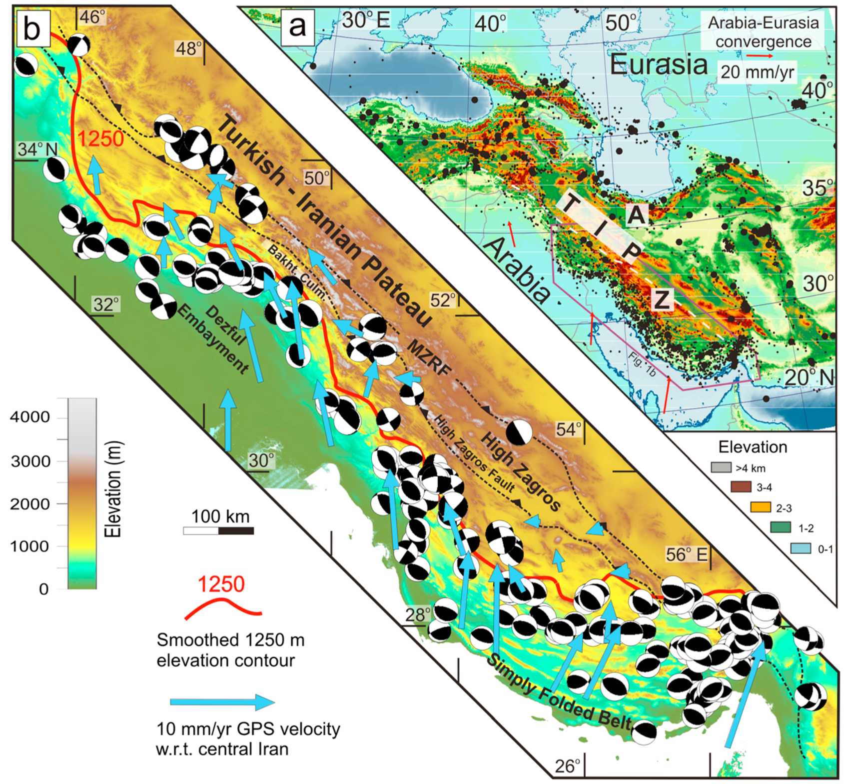

Figure 1. (a) Regional topography and seismicity of the Arabia-Eurasia collision. Large dots are epicenters of earthquakes of $M > 6$ from 1900 to 2000 [*Jackson*, 2001], small dots are epicenters from the EHB catalogue 1964–1999, $M > \sim 5$. Red arrows show GPS-derived velocity with respect to Asia from *Sella et al.* [2002]. A = Alborz; TIP = Turkish-Iranian plateau; Z = Zagros. (b) Seismicity of the Zagros: focal mechanisms reported in *Nissen et al.* [2011] and references therein. Note the scarcity of thrusts above the smoothed 1250 m regional elevation contour (derived using a Gaussian filter with a radius of 50 km). Earthquake epicenters are accurate to within 20 km [*Nissen et al.*, 2011]. GPS vectors are from *Walpersdorf et al.* [2006]. MZRF = Main Zagros Reverse Fault (Zagros suture).

seismicity and topography is that it is harder to continue to shorten crust that has already been significantly thickened and elevated than it is to shorten the thinner crust toward the foreland [e.g., *Dalmayrac and Molnar*, 1981]. Shear stresses are likely to be highest near gradients in crustal thickness [*England and McKenzie*, 1982], which has been invoked as the reason for the distribution of Zagros earthquakes [*Jackson and McKenzie*, 1984].

[5] What is not clear is whether such plateaux grow incrementally or in discrete re-organizations [*Tapponnier et al.*, 2001]. Evidence is emerging that early deformation in collision zones can occur thousands of kilometers from the original suture [*Vincent et al.*, 2007; *Clark et al.*, 2010], but this is not the same as determining when compressional deformation and surface uplift cease in each area, i.e., the transition from an active fold-and-thrust belt into an orogenic plateau. Nor is it understood whether they grow by achieving a critical thickness and elevation first in a limited area, which then laterally enlarges [*Rowley and Currie*, 2006], or by a widespread rise in surface elevation to present values. Such a rise could be protracted [*Barnes and Ehlers*, 2009], or sudden, possibly as the result of loss of the lower lithosphere [*Garzione et al.*, 2006].

[6] Lithospheric thickness estimates for Asia based on shear wave velocity gradients [*Priestley and McKenzie*, 2006] reveal a core near the original Arabia-Eurasia suture with thicknesses over 200 km, thinning to near normal (100–120 km) values farther north and south. See also *Kaviani et al.* [2007].

[7] Our approach to understanding the growth of the Turkish-Iranian Plateau is to combine existing seismicity and GPS data (which provide the kinematic framework for the study area) with new observations on the geology and geomorphology. Sections 2.1 to 2.3 summarize the geological background of the Arabia-Eurasia collision and the Zagros range in particular. We then provide a summary of the seismotectonics and GPS-derived velocity field, as these provide essential constraints on the location of active deformation. We utilize the seismicity data in constructing two new crustal-scale cross-sections through the Zagros, based on published geological maps, available sub-surface data, satellite imagery, and our fieldwork observations (section 3). These cross-sections emphasize the along-strike variation in the structure of the Zagros, including (1) the nature of the Dezful Embayment (Figure 2) and (2) the lower degree of shortening in the High Zagros in the southeast. We examine the geomorphology of the Zagros and the adjacent part of the Turkish-Iranian Plateau in section 4, as the tectonics of a region may be recorded in its landscape. Digital elevation models, satellite imagery, and our fieldwork observations show the relationships of the landscape to the deep structure and the seismicity.

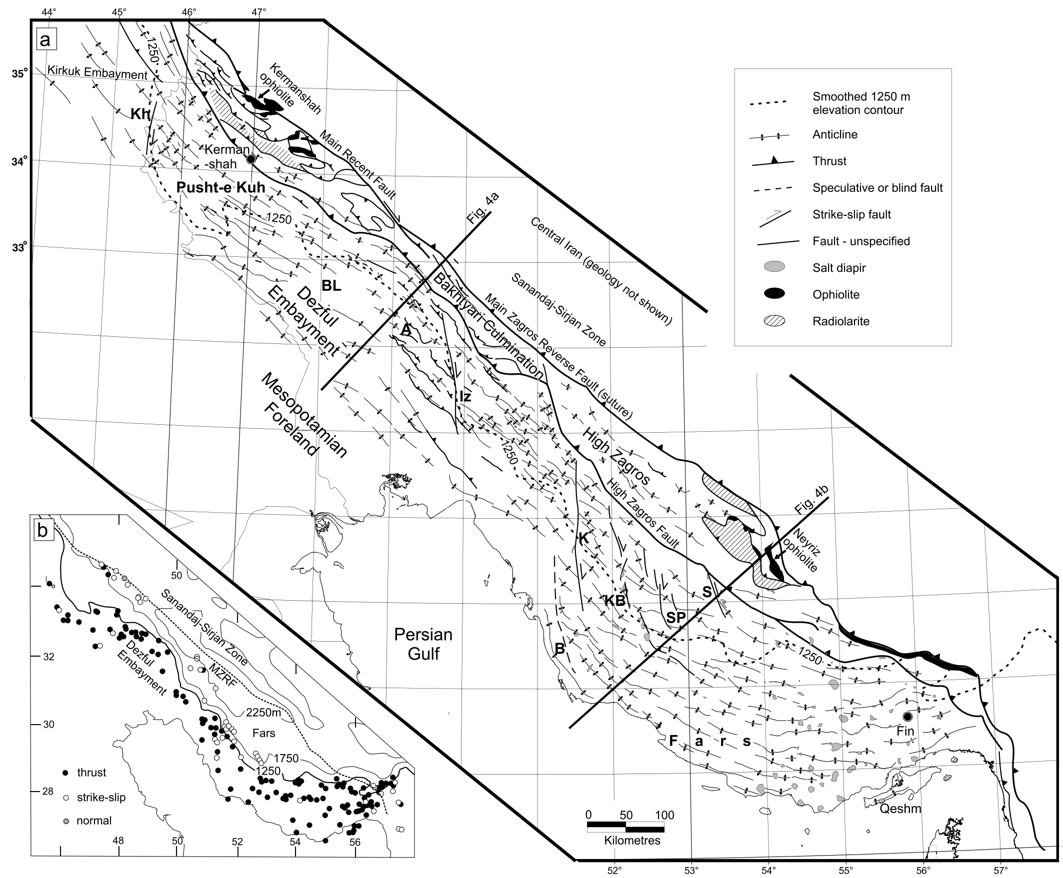

Figure 2. (a) Location map and major structures of the Zagros Simply Folded Belt, Iran. Derived from *NIOC* [1975, 1977], *Berberian* [1995], *Hessami et al.* [2001], *Blanc et al.* [2003], *Agard et al.* [2005], and *Babaie et al.* [2006]. Key to fault abbreviations: B = Borazjan; Iz = Izeh; K = Kazerun; KB = Kareh Bas; Kh = Khanaqin; S = Sarvestan; SP = Sabz-Pushan; BL = Balarud Line; A = Kuh-e Asmari. b) Earthquake epicentres across the Zagros, from *Nissen et al.* [2011] and references therein, divided by fault type. MZRF = Main Zagros Reverse Fault.

2. Regional Tectonics

2.1. Collision Overview

[8] The Bitlis-Zagros suture between the Arabian and Eurasian plates follows a convex-northward line within SE Turkey and runs NW-SE through northern Iraq and southern Iran (Figure 1). South and southwest of the suture, the original passive continental margin of the Arabian plate is now deformed by folds, thrusts, and strike-slip faults within the Zagros mountains [e.g., *Jackson and McKenzie*, 1984; *Alavi*, 1994; *Berberian*, 1995; *Agard et al.*, 2011; *Mouthereau et al.*, 2012]. The present day deformation front lies approximately parallel to the Iranian shoreline of the Persian Gulf; Cenozoic folds occur within the northeastern Gulf, detected on sub-surface seismic data [*Soleimany and Sabat*, 2010]. The Gulf and the neighboring plains of Mesopotamia represent the flexural foredeep to the Zagros fold-and-thrust belt [*Beydoun et al.*, 1992; *Aqrawi et al.*, 2010]. The original Eurasian margin lies along the southwest side of the Sanandaj-Sirjan Zone, which is an elongate block of crust that accreted to other blocks of crust within Iran during the Cretaceous [*Şengör et al.*, 1988].

[9] The timing of initial Arabia-Eurasia collision is still debated, with recently published estimates spanning from the Late Cretaceous/Paleocene [*Mazhari et al.*, 2009], the Early Miocene [*Okay et al.*, 2010] to the Mid/Late Miocene [*Guest et al.*, 2006]. *Allen and Armstrong* [2008] summarized geological evidence from both sides of the suture that indicate a Late Eocene (~35 Ma) age for initial collision, including a sharp reduction in magmatism and the development of unconformities on both the Arabian and Eurasian plates. *Ballato et al.* [2011] and *Mouthereau et al.* [2012] made the case for initial collision at ~35 Ma followed by intensification of deformation at ~20 Ma, perhaps caused by the end of underthrusting of thin Arabian margin crust under Eurasia, and so the onset of deformation of the interior of the Arabian plate [*Morley et al.*, 2009].

[10] This intensification is becoming more clear from a combination of exhumation and sedimentology/provenance studies [e.g., *Fakhiri et al.*, 2008; *Khadivi et al.*, 2010; 2012; *Gavillot et al.*, 2010; *Rezaeian et al.*, 2012], but it is not clear whether there was a smooth, incremental migration of deformation and, potentially, plateau development toward the foreland, or if there was a more discontinuous progression in a series of abrupt re-organizations.

[11] There was possibly also a re-organization of the collision zone at ~5 Ma [e.g., *Axen et al.*, 2001]; many of the active fault systems within the collision zone seem to

have originated or intensified at this time [*Allen et al*. 2004]. There was no abrupt change in Arabia-Eurasia convergence rates at ~5 Ma; if anything, convergence has slowed in the last couple of million years [*DeMets et al*., 1994; *Sella et al*., 2002]. Therefore, the re-organization may relate to a different cause, such as break-off of the Neo-Tethyan oceanic slab [*Keskin*, 2003] or the closure of the free face at the eastern margin of the collision zone [*Allen et al*., 2011a].

[12] The Arabia-Eurasia collision actively deforms much of SW Asia between western Turkey and eastern Iran, across an area of ~3×10^6 km^2 [*Allen et al*., 2004; *Vernant et al*., 2004; *Agard et al*., 2011; *Mouthereau*, 2011; *Mouthereau et al*., 2012]. Deformation is distributed across a broad region from the Persian Gulf to the Caucasus, Alborz and Kopet Dagh ranges, and between the Aegean and eastern Iran. Seismicity is concentrated at fold-and-thrust belts at the margins of this region, particularly the Zagros in the south and the Greater Caucasus-Alborz-Kopet Dagh ranges in the north [Figure 1; *Jackson and McKenzie* 1984; *Talebian and Jackson*, 2004].

[13] The Turkish-Iranian Plateau covers ~1.5×10^6 km^2, mostly within the territory of Iran and Turkey (Figure 1). Most of the plateau area belonged to the Eurasian plate before collision with Arabia. Elevations are typically over 1.5 km, with subdued relief compared with the ranges to its north and south. The geomorphic expression of the plateau corresponds roughly with a division in the active tectonics: the precise correspondence is described later in this paper. Much of the plateau interior is seismically inactive, or affected by strike-slip faults that have a variety of kinematic roles including strain partitioning with fold-and-thrust belts, collision zone boundaries, and shortening by vertical axis rotations [*Talebian and Jackson*, 2002; *Walker et al*., 2004; *Meyer et al*., 2006; *Dhont et al*., 2006; *Allen et al*., 2011a].

2.2. Zagros Stratigraphy

[14] Stratigraphy in the Zagros records the evolution from the passive margin of the Arabian plate to the foreland basin of the Arabia-Eurasia collision zone (Figure 3), although as noted above, the precise time of the onset of continental collision is debated. Precambrian basement is not exposed in situ but is recorded as blocks brought to the surface in salt diapirs [*Kent*, 1979], derived from the Hormuz Series. The Hormuz Series is of late Precambrian/Cambrian age and overlies the basement. The Series is present across a wide area of the Zagros and Middle East [*Edgell*, 1991] and contains thick evaporites, mainly halite. The original distribution of these evaporites is not known. Some geologists [e.g., *Murris*, 1980; *Edgell*, 1991] infer that the present distribution of diapirs is a guide to the original depositional extent: these occur across large areas of the eastern Zagros (Fars), but not within the Dezful Embayment or farther west (Figure 1). Hormuz Series salt also crops out in the High Zagros to the north of the Dezful Embayment. Areas without salt exposures were regions of clastic deposition or non-deposition in such schemes. Other studies extend the distribution of evaporites throughout the Zagros, with later factors controlling the present distribution of diapirs [e.g., *Kent*, 1979].

[15] The Hormuz Series lies at the base of a Palaeozoic platform succession, similar to strata deposited over Pan-African basement across North Africa and Arabia [*Holt et al*., 2010].

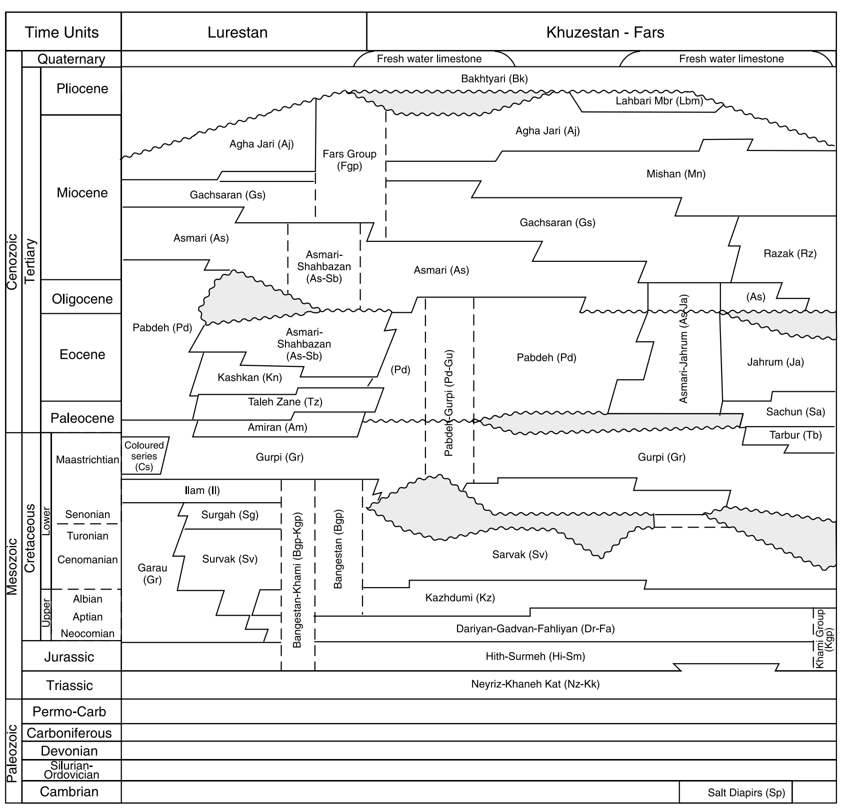

Figure 3. Stratigraphy of the Iranian Zagros. Modified from *Iran Oil Operating Companies* [1969] to reflect the diachronous nature of the Bakhtyari Formation [*Fakhari et al*., 2008].

This is inferred to be interrupted by Permian rifting, which led to the spreading of the Neo-Tethyan Ocean from the contemporary margin of Gondwana [*Şengör et al.*, 1988], although the evidence on which this is based remains sparse [*Szabo and Kheradpir*, 1978]. *Sepehr and Cosgrove* [2004] illustrate an example of a Permian-Triassic half graben from the Dezful Embayment. A ~4–5 km thick Mesozoic and lower Tertiary succession contains alternating deposits of clastics and carbonates [*Setudhina*, 1978, *Sherkati and Letouzey*, 2004]. Mid Cretaceous (Bangestan Group) and Oligo-Miocene (Asmari) limestones form prominent topographic markers in the Zagros landscape, by virtue of their high mechanical strength and consequent low erosion rates.

[16] Gachsaran Formation evaporites (Miocene) form a second mobile unit in the stratigraphy, below a Miocene-Quaternary clastic succession that coarsens upward [*Homke et al.*, 2004]. Classical stratigraphies for the Zagros indicate a pulse of coarse conglomerates above an intra-Pliocene unconformity [*James and Wynd*, 1965]. Recent work shows that these conglomeratic units are highly diachronous across the range [*Fakhari et al.*, 2008; *Khadivi et al.*, 2010] and most likely represent the switch to locally derived sediments in each area as deformation and relief proceed across the fold-and-thrust belt [*Pirouz et al.*, 2011].

2.3. Zagros Tectonics

[17] The simplest tectonic scheme for the Zagros has two divisions (Figure 2), southwest of the suture that is known as the Main Zagros Reverse Fault. The High Zagros (or Imbricate Belt) lies between the original suture and a major thrust, the High Zagros Fault, which is mapped as running roughly parallel to the suture and 40–160 km southwest of it [*Alavi*, 1994; *Berberian*, 1995; *Bosold et al.*, 2005]. The remainder of the Zagros is the Simply Folded Belt.

[18] The High Zagros is commonly depicted as a single tectonic unit along the length of the range, but it is not [see, for example, *Authemayou et al.*, 2006]. Nor does the High Zagros Fault have a similar character along the length of the range. In the northwest, the High Zagros consists of low angle nappes that contain the following groups of lithologies: (i) deepwater Triassic-Cretaceous clastics known as the Radiolarite Group, (ii) Triassic-Cretaceous limestones of the Bisotun Limestone, (iii) the Kermanshah ophiolite with Late Cretaceous generation and emplacement ages, (iv) low grade metamorphics (phyllites) derived from Triassic-Cretaceous sediments, (v) Barremian-Aptian Orbitolina Limestones, and (vi) Eocene volcanics [*Braud*, 1970; *Ghazi and Hassanipak*, 1999; *Agard et al.*, 2005]. The last three sets of lithologies derive from the Sanandaj-Sirjan Zone, i.e., on the Eurasian side of the suture, and are overthrust toward the southwest in one main sheet subdivided into three units [*Agard et al.*, 2005]. The Arabia-Eurasia suture *sensu stricto* (i.e., Main Zagros Reverse Fault) lies underneath the lowermost unit derived from the Sanandaj-Sirjan Zone. Upper Oligocene-Lower Miocene conglomerates and limestones lie unconformably over the nappes and are themselves overlain by Miocene flysch [*Agard et al.*, 2005]. The High Zagros Fault is mapped as a low angle thrust at the base of a nappe containing the Radiolarite Group [*National Iranian Oil Company* (*NIOC*), 1975].

[19] Northeast of the Dezful Embayment, the higher ground of the Simply Folded Belt and the High Zagros are known collectively as the Bakhtyari Culmination. Highly imbricated slices of Palaeozoic and Mesozoic strata within the High Zagros [*Nemati and Yassaghi*, 2010] are inferred to originate on the Arabian plate. There are outcrops of the Hormuz Series salt along fault planes. Here there are the highest summits (>4000 m) and deepest erosion levels of the Zagros; rocks as old as the Cambrian are locally exposed, and the regional exposure level is typically at the Cretaceous Bangestan Group. The High Zagros Fault in this area is interpreted to cut the basement [*Bosold et al.*, 2005; *Authemayou et al.*, 2006] and overthrust to the southwest.

[20] The Main Zagros Reverse Fault is more linear than to the northwest and juxtaposes the Palaeozoic-Cretaceous stratigraphy against the low grade metamorphics and Cretaceous cover sequence of the Sanandaj-Sirjan Zone [*Authemayou et al.*, 2006]. Along strike to the southeast, within the Fars region, folds within the High Zagros become more open, indicating that the structure (and percentage shortening?) of the High Zagros is not the same along strike [*NIOC*, 1975; 1977]. Exposures of Palaeozoic strata and exposed thrusts both become rarer to the southeast. The commonest unit exposed is the Bangestan Group, but outliers preserve Tertiary strata. The High Zagros Fault is mapped at the southwestern regional limit of the Bangestan Group but is not typically an exposed fault [*Berberian*, 1995]. At the northeastern side of the High Zagros in Fars, there is the Neyriz ophiolite and associated melanges (Figure 2), thrust to the southwest over the Arabian plate stratigraphy [*Babaie et al.*, 2006]. These units are the equivalents of the Kermanshah ophiolite along strike to the northwest. They are adjacent to the Main Zagros Reverse Fault in this area, which is mapped as a steep thrust fault. There are no known low angle nappes derived from the Sanandaj-Sirjan Zone.

[21] The greater part of the Zagros lies within the Simply Folded Belt, where spectacular "whaleback" anticlines deform the sedimentary cover of the Arabian plate and expose the Mesozoic-Cenozoic mixed carbonate-clastic succession. Palaeozoic strata are very rarely exposed. The Hormuz Series salt crops out in diapirs in the east of the Zagros [*Gansser* 1992; *Talbot and Alavi*, 1996], east of the Kazerun Fault in the Fars region (Figure 2). Within Fars, there is a triangular zone tapering northward where there are no salt diapirs (Figure 2).

[22] The regional southern limit of Asmari Limestone exposure forms a distinct topographic front within the Zagros, referred to as the Mountain Front [*McQuillan*, 1991]. It has been interpreted as marking the position of a major, albeit segmented thrust front that runs along most of the Zagros: the Mountain Front Fault [*Berberian*, 1995; *Casciello et al.*, 2009]. In the southeastern Zagros, it splays off another discontinuous series of faults for which *Berberian* [1995] used the term "Zagros Foredeep Fault," to refer to a blind "master" thrust that causes uplift along the Zagros foreland.

[23] The Zagros has two embayment features, the Dezful Embayment in Iran, and the similar but less well known Kirkuk (Kirkut, Karkuk) Embayment in Iraq [Figure 2; *Carruba et al.*, 2006; *Aqrawi et al.*, 2010]. These are defined by subdued relief and exhumation compared with zones along strike to the NW and SE. Thus, the Asmari Limestone is not exposed within the Dezful Embayment, except at Kuh-e Asmari, but crops out all around it. The origin of these

embayments is not certain. Superficially, they resemble re-entrants in the front of a propagating thrust belt [*Bahroudi and Koyi*, 2003], but the Dezful Embayment does not represent a significant change in the position of the frontal structures of the Zagros: Compressional structures are present within it, and the position of the deformation front is roughly linear along the southwestern margin of the range (Figure 2). The high ground between the Dezful and Kirkuk embayments is known as the Pusht-e Kuh arc or the Lurestan arc.

[24] Accommodation of roughly north-south plate convergence is achieved by a combination of thrust and strike-slip faults across the Zagros, in an example of strain partitioning. Right-lateral strike-slip along the Main Recent Fault transfers to the southeast into a diverging array of faults [*Talebian and Jackson*, 2002; *Authemayou et al*., 2009].

[25] There are now many estimates for shortening across the Zagros, based on balanced cross-sections. Some estimates put northeast-southwest shortening across the Simply Folded Belt in the order of 40–60 km, and roughly consistent whichever transect across the range is used [e.g., *Blanc et al*., 2003; *McQuarrie*, 2004; *Alavi*, 2007]. There are some lower estimates that interpret very little faulting in the Zagros cover sequence above the Hormuz Series: *Sherkati and Letouzey* [2004] and *Mouthereau et al*. [2007] calculated 21 and 15 km of shortening across Fars, while *Vergés et al*. [2011] calculated 21 km for a transect across the Pusht-e Kuh arc. Most estimates that include the High Zagros only add ~20 km to these totals [e.g., *McQuarrie*, 2004].

[26] Fewer estimates have been made for the amount of underthrusting of Arabia beneath the Eurasian plate, deeper-level shortening of this underthrust crust, or shortening within Eurasian units thrust over the Arabian plate. *Vergés et al*. [2011] used area balancing techniques to suggest a total of 149–180 km shortening across the Arabian plate. *Mouthereau* [2011] deduced 135 km of total Arabian plate shortening, by summing earlier estimates for different parts of the Zagros. A portion of the thin, leading edge of the Arabian plate has apparently underthrust Eurasia and is detected on deep seismic profiles across Iran [*Paul et al*., 2010], as far as 170–270 km north of the surface trace of the suture. This range of values is distinctly higher than the 50 km of underthrusting calculated by *Mouthereau et al*. [2007] from structural restorations. *Agard et al*. [2005] found ~70 km of shortening adjacent to the suture in the northwest Zagros, where units originating in the Sanandaj-Sirjan Zone overthrust the Arabia margin.

[27] Teleseismic receiver function analysis provides information on crustal thickness under the Zagros, although the precise variation is debated. *Paul et al*. [2010] showed that crustal thickness is roughly constant across much of the Zagros at 42 ± 2 km (moving southwest to northeast across the strike of the range, starting from the undeformed foreland of the Persian Gulf region), to within a few 10s of kilometers southwest of the Zagros suture. Farther northeast, the crustal thickness increases to 55–70 km, with the maximum close to the line of the suture or the southwest part of the Sanandaj-Sirjan Zone immediately to its northeast. *Rham* [2009] described a more steady increase in Moho depth, from ~40 km near the southwest limit of the Zagros to ~55 km at the suture. The Arabian plate southwest of the Zagros has a crustal thickness of ~40 km [*Gok et al*., 2008].

2.4. Seismicity and Geodesy

[28] The northern promontory of the Arabian plate presently converges with Eurasia at a rate of ~15–18 mm yr^{-1}, in an approximately NNW direction [*McClusky et al*. 2000]. Convergence rates increase eastward, because the Arabia-Iran pole of rotation lies within the eastern Mediterranean region [*Jackson and McKenzie*, 1988] and is roughly 26 mm yr^{-1} at the longitude of eastern Iran [*Sella et al*. 2002] (Figure 1). These estimates are ~10 mm yr^{-1} lower than the NUVEL-1A plate motion model of *DeMets et al*. [1994].

[29] Global Positioning System campaigns provide information about the distribution of active strain within the Zagros [*Vernant et al*., 2004; *Hessami et al*., 2006; *Walpersdorf et al*., 2006]. Active north-south convergence across the Zagros takes place at up to ~10 mm yr^{-1}, but this is concentrated in the lower elevation parts of the range [*Oveisi et al*., 2009] and decreases northward toward the Turkish-Iranian Plateau. Internal deformation within the plateau, north of the Zagros, takes place at less than 2 mm yr^{-1}, which is roughly 10% of the overall Arabia-Eurasia convergence rate [*Vernant et al*., 2004].

[30] Seismicity in the Zagros is concentrated between the deformation front in the foreland and the regional 1250 m contour [Figure 1; *Jackson and McKenzie*, 1984; *Talebian and Jackson*, 2004; *Nissen et al*., 2011]. The fit is even better if only thrust events are considered, because it leaves out strike-slip events along the Main Recent Fault and along several of the right-lateral faults to the east of the Dezful Embayment.

[31] Earthquake depths provide valuable information about the style of local deformation and the strength profile of the lithosphere. Zagros earthquake studies confirm that the crystalline basement deforms at up to depths of 20 km with earthquakes of M ~5–6, but no deeper [*Baker et al*., 1993; *Maggi et al*., 2000; *Talebian and Jackson*, 2004; *Tatar et al*., 2004; *Adams et al*., 2009; *Nissen et al*., 2011]. There is some deeper microseismicity in a few areas, such as at Fin (Figure 2), where there is an abundance of aftershocks at 20–25 km and a few as deep as 30 km, and near Kermanshah (Figure 2), where there are also a few well located events at ~30 km [*Nissen et al*., 2011]. Note there is a rough equivalence between the uncertainty in the hypocenter depth ($\pm$ 4km) and the down-dip length of fault ruptured in events of M ~5–6, i.e., even if such an earthquake really occurred at 4 km above its quoted depth, the fault would have ruptured to that depth, given the magnitude of the event. Some earthquakes must occur within the sedimentary cover [*Adams et al*., 2009], given that their depths are >4 km above the likely depth to basement. *Roustaei et al*. [2010] and *Nissen et al*. [2007, 2010] suggested that the sequences of M ~5–6 earthquakes in Fin (in 2006) and Qeshm (in 2005–2009), both in the southeast Simply Folded Belt, were generated entirely within sedimentary cover. At least some of these events involved thrust faults that dip southward, toward the foreland. *Nissen et al*. [2011] used earthquake depth and magnitude distributions across the Zagros to propose that ruptures of M ~5 events typically affected either the sedimentary cover or the basement, but not both, consistent with the Hormuz Series salt forming an effective barrier to rupture propagation at the base of the sedimentary succession. Rarer earthquakes of M ~6.5 produce ruptures so large they must cut across the salt. These earthquake data have guided our structural interpretation in Figure 4.

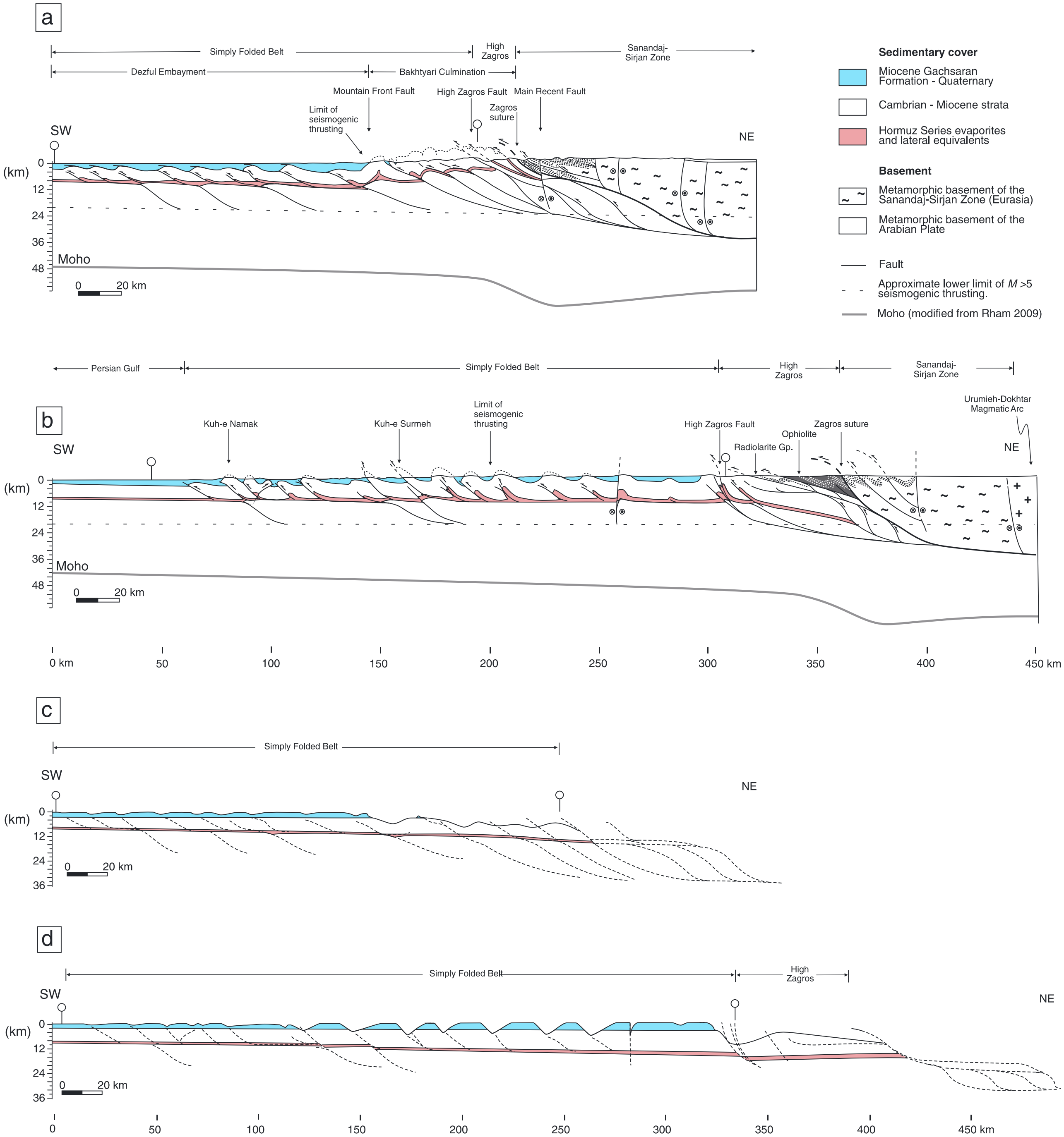

Figure 4. (a) Cross-section through the Dezful Embayment and the Bakhtyari Culmination. (b) Cross-section through Fars. Locations shown on Figure 2. (c and d) Restored and balanced sections for Figures 4a and 4b. Data are derived from SRTM topography [*Jarvis et al*. 2008], NIOC 1:1,000,000 maps [1975, 1977], 1:250,000 maps of *Llewellyn* [1972, 1973], and satellite imagery from GoogleEarth. Although it is likely that most basement faults dip to the north, as shown, this is not conclusive in the seismicity data.

[32] Low angle seismogenic thrusts are rare [*Talebian and Jackson*, 2004; *Nissen et al*., 2011]. This does not rule out aseismic slip along detachments, which would be more likely along the weak sedimentary units that act effectively as detachments in the first place.

[33] The GPS-derived velocity field and the seismicity record represent different aspects of the strain record across the region. That they both show decadal-scale quiescence across the higher elevation parts of the Zagros confirms the absence of active upper crustal shortening within this region.

3. Structural Geology: Regional Cross-sections

[34] Figures 4a and 4b are cross-sections through the Zagros, in the Dezful Embayment/Bakhtyari Culmination and Fars regions, respectively. They are based on

combinations of 1:100,000 and 1:250,000 geological maps, satellite image interpretation, seismicity, fieldwork within and at the margins of the Dezful Embayment, and reconnaissance fieldwork in the Fars region. Debate continues as to the sub-surface structure of the Zagros [*Casciello et al*., 2009; *Farzipour-Saein et al*., 2009; *Vergés et al*., 2011; *Yamato et al*., 2011]. The problem is that there are few sub-surface data, especially for depths >10 km where most controversy exists over fold and fault geometry. It is possible to draw fold and fault geometries that are very different at depth, based on the same surface and near-surface data. One end-member model emphasizes detachment of the cover from the basement, particularly along the Hormuz Series salt, but also on lateral equivalents where the evaporites are not known to exist [*Stöcklin*, 1968; *McQuarrie*, 2004; *Yamato et al*., 2011]. The other end-member links each exposed fold to a basement thrust [*Ameen*, 1991]. It is entirely feasible that both thick-skinned and thin-skinned styles of deformation occur in the range, with multiple detachments within the sedimentary cover, co-eval with basement thrusts [*Blanc et al*., 2003; *Ahmadhadi et al*., 2007; *Casciello et al*., 2009; *Mouthereau et al*., 2006; 2007]. The latter are especially likely at zones of high structural relief, the so-called master faults of the Zagros [*Berberian*, 1995].

[35] Seismicity clearly indicates that the basement actively deforms with events of *M* ~5-6 [*Baker et al*., 1993; *Talebian and Jackson*, 2004; *Tatar et al*., 2004; *Nissen et al*., 2011]: Hypocenters of up to 20 km depth are located beneath the sedimentary cover, even allowing for uncertainty in the precise depth of the hypocenter and the thickness of the stratigraphy. We use this constraint to guide our interpretation of Figure 4, locating a major basement thrust at features with particular structural relief, such as the Mountain Front Fault [*Berberian*, 1995]. Other basement thrusts are likely, given the seismicity record, but it is hard to know their exact relation to exposed folds; the locations of epicenters are not precise enough.

[36] There are also sufficient data to show that most (~75%) earthquakes of *M* ~5–6 are located within the cover and do not cut across the Hormuz Series salt or any lateral equivalents [*Nissen et al*., 2011]. This constraint is also honored in the cross-sections: Thrusts are blind and may not cut through carbonate anticlines exposed at the surface, but they do exist under the Zagros Simply Folded Belt within the cover stratigraphy.

3.1. Dezful Embayment/Bakhtyari Culmination

[37] The section in Figure 4a covers the Dezful Embayment and higher terrain in the Bakhtyari Culmination to its northeast. The Dezful Embayment is a low relief, low elevation region within the Simply Folded Belt, covering ~75,000 km^2. The origin of the Embayment may relate to the pre-collision structural evolution of the Zagros, and the discontinuous distribution of ophiolite nappes along its northeast margin: These are lacking adjacent to the Embayment, in contrast to the regions to the north and southeast [*Allen and Talebian*, 2011]. The Cenozoic sedimentary succession reaches over 5 km in thickness [*Koop and Stoneley*, 1982], and alluvial deposition continues across much of the Embayment at present. However, it is not internally undeformed but contains anticlines that act as hydrocarbon traps to Iran's most important hydrocarbon fields [*Beydoun et al*., 1992; *Carruba et al*., 2006]. These include the Ahwaz anticline at the southwestern side of the Embayment (Figure 2). Seismicity occurs within the Embayment [*Talebian and Jackson*, 2004], and the fold geomorphology indicates active growth [*Allen and Talebian*, 2011], reinforcing the point about active deformation. The folds within the Dezful Embayment typically do not expose the Asmari Limestone or older units (Figure 4a), permitting the hydrocarbon systems to operate without breaching of the traps.

[38] Like most previous cross-sections for the Zagros [e.g., *Blanc et al*., 2003; *McQuarrie*, 2004, *Sherkati et al*., 2006; *Farzipour-Saein et al*., 2009], we show little or no basement relief where there is no change in structural elevation on either side of anticlines. This is the simplest reconstruction but is an over-simplification given that the regional structure is affected by rifting during the Permian-Triassic [*Sepehr and Cosgrove*, 2004] and probably older and younger intervals [*Frizon de Lamotte et al*. 2011]. In such cases, inversion of rift faults would involve the basement beneath anticlines but not produce significant structural relief at higher, exposed, levels of the stratigraphy.

[39] Our interpretation depicts the folds as being strongly linked to underlying thrusts. This view is strongly influenced by the seismicity record: It is hard to see how a region affected by numerous *M* ~5–6 events cannot be affected by fault slip (recalling that an event of *M* ~6 produces slip in the order of 1 m). Several papers model the Zagros folds as buckle structures (e.g., *Yamato et al*., 2011), but this approach does not account for the seismicity.

[40] The Gachsaran evaporites form a detachment between overlying and underlying stratigraphy, and mean that the uppermost parts of folds can be disharmonic with the underlying structure [*O'Brien*, 1957; *Sepehr et al*., 2006]. Brittle thrusts deform this unit where it crops out on the northeast side of the Dezful Embayment (Figure 5a) and place it over younger strata of the Agha Jari Formation. Such faults are depicted schematically on Figure 4a, which does not attempt to represent the detailed structure within and above the Gachsaran Formation. Other plausible décollement units include the Cretaceous Kazhdumi Formation (mudrocks) and the Triassic Dashtak Formation (evaporites) [*Sepehr et al*., 2006; *Vergés et al*., 2011].

[41] An important structural step in this section line comes at the Mountain Front Fault (Figure 5b), where the exposure level deepens to involve the Asmari Limestone, suggesting a regional elevation of basement on the kilometer scale [*Berberian*, 1995]. Anticlines to the northeast, within the remaining ~30 km of the Simply Folded Belt, expose the Bangestan Group. Major anticlines within the Simply Folded Belt are not adjacent to exposed thrusts but are interpreted to overlie blind thrusts (Figure 4a).

[42] The High Zagros Fault is the southwestern exposed thrust associated with a major hanging wall anticline [*Bosold et al*., 2005] and marks the boundary of the High Zagros. The High Zagros is at its narrowest where it is northeast of the Dezful Embayment (Bakhtyari Culmination), with a width of ≤50 km. This contrasts with the Fars region, where it is mapped as 100 km wide. There is much more imbrication in the High Zagros, compared with the Simply Folded Belt within and adjacent to the Dezful Embayment [*NIOC*, 1975; *Authemayou et al*., 2006; *Nemati and Yassaghi*, 2010]. Structural vergence is generally toward the southwest

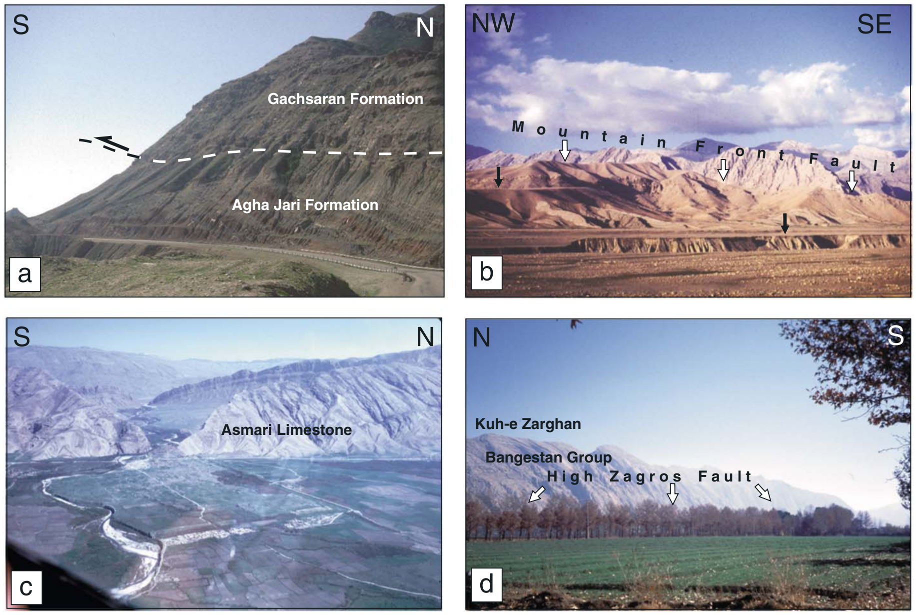

Figure 5. Field photographs of Zagros structures. (a) Gachsaran Formation thrust southwest over Agha Jari Formation clastics at the northeast side of the Dezful Embayment. Photo taken from ~32.03°N 49.13°E. (b) View north of the Mountain Front Fault at Kuh-e Kamar Meh, taken from ~32.47°N 49.15°E. Solid arrows mark terraces within Quaternary river gravels; open arrows mark the trace of the Mountain Front Fault. (c) View west from a helicopter of Tang-e Chowgan (Shahpur), taken from 29.8°N 51.65°E. Photograph courtesy of Peter Llewellyn. (d) View east of Kuh-e Zarghan, along the High Zagros Fault, taken from ~29.78°N 52.71°E.

in the High Zagros; thrusts dip northeast. The presence of Hormuz Series salt along thrust planes and the lack of basement exposures suggest that the salt strongly decouples the cover and basement in this area. The spacing of thrusts decreases toward the Main Recent Fault, which is an active, right-lateral strike-slip fault [*Talebian and Jackson*, 2002; *Authemayou et al*., 2009], roughly along the line of the original suture between the Arabian and Eurasian plates (Main Zagros Reverse Fault). We depict the Main Recent Fault as dipping steeply, and cross-cutting inactive thrusts of the High Zagros at depth, based largely on the seismicity pattern in the area [*Nissen et al*., 2011].

[43] We find no evidence for two separate structural styles or events in the folds, reported in the far east of the Zagros as evidence for a change from thin-skinned to thick-skinned thrusting over time [*Molinaro et al*., 2005]. Therefore, such a two-stage structural evolution is not considered further in our interpretation of the main part of the Zagros.

[44] The northeastern limit of significant seismogenic thrusting in this region lies within the Simply Folded Belt, to the southwest of the High Zagros Fault. There is no single structure associated with the cutoff, although earthquakes are concentrated in the region where topography climbs from the plains of the Dezful Embayment but elevations are below 750 m.

[45] Figure 4c is a restored section for the line of Figure 4a. The Simply Folded Belt is estimated to have shortened by ~55 km or roughly 20%. This is line with previous estimates for the same region [e.g., *Blanc et al*., 2003; *McQuarrie*, 2004]. Shortening for the entire Arabian plate is harder to constrain, because of the highly deformed nature of the High Zagros and the presumed deformation of rocks underthrust below the suture zone. We tentatively estimate shortening for the Arabian plate as far as the suture as ~110 km. Underthrusting of the Arabian plate beneath Eurasia and shortening within Eurasia need to be added to this value to calculate total convergence in the collision zone.

3.2. Fars

[46] The structure across the Fars section line (Figure 4b) is much more uniform than the Dezful/Bakhtyari Culmination transect, without an equivalent division between a region of low relief/elevation/exhumation and a corresponding imbricate zone to the northeast [*Mouthereau et al*., 2007]. Although some equivalent structural units and boundaries have been defined and mapped [*Berberian*, 1995], they have little in common with the Dezful/Bakhtyari Culmination region depicted in Figure 4a. The Mountain Front Fault is usually mapped at the southernmost exposure of Asmari Limestone, which is within 20 km of the modern shoreline of the Persian Gulf (Kuh-e Namak on Figure 4b). The most dissected anticlines along this line expose Cretaceous limestones of the Bangestan Group. Some authors place the Mountain Front Fault ~100 km further north, along a line of anticlines that also expose the Bangestan Group [e.g., *Sepehr and Cosgrove*, 2004]. This is at Kuh-e Surmeh along the section line of Figure 4b. There is no regional equivalent to the subdued folds and relief of the Dezful Embayment. *Berberian* [1995] used this geomorphic marker to suggest that the Mountain Front Fault was originally continuous along the Zagros and is offset by 10s of kilometers along the Kazerun Fault, at the eastern side of the Dezful Embayment. However, detailed structural studies along the Kazerun Fault reveal much

smaller offsets, on the order of 10 km [*Authemayou et al.*, 2006; *Lacombe et al.*, 2006]. Therefore, the southernmost appearance of the Asmari Limestone is a prominent geomorphic marker along the Zagros, but not one that marks an original, contiguous fault zone. The term “Mountain Front Fault” needs to be used with this caveat in mind.

[47] Other folds within the Simply Folded Belt to the north also expose the Asmari Limestone, and locally the Bangestan Group. These are classic Zagros whaleback anticlines, with half wavelengths in the order of 10 km and slight vergence toward the south (Figure 5c). As for the Dezful region, the deep structure of these folds is not clear, nor how they relate to underlying detachment(s) in the sedimentary cover and thrusts in the basement. *Tatar et al.* [2004] concluded that there need not be a one-to-one correlation between basement faults and observed folds, based on the spacing between seismic lineaments of 15–20 km and a calculated fold spacing of 10–15 km. (However, *Mouthereau et al.* [2007] reported an average wavelength of ~16 ± 5 km for 149 Zagros folds, essentially the same as this seismic lineament spacing.) Some thrusts may dip to the south, where there is little or no asymmetry in the overlying folds [*Roustaei et al.*, 2010]. There is no obvious change in fold style northeast of the limit of seismogenic thrusting (Figure 4b).

[48] The High Zagros Fault is conventionally mapped within Fars as a thrust at the southern limit of the regional exposure of the Bangestan Group [*Berberian*, 1995]. The distinction is not completely clear-cut. Lower Tertiary strata crop out, albeit locally, northeast of the defined High Zagros Fault trace [e.g., *NIOC*, 1977], and as noted above, there are Bangestan Group outliers in the cores of many of the anticlines within the Simply Folded Belt. Nor is a thrust exposed in the region of the cross-section: The High Zagros Fault is presumed to be blind. Our cross-section (Figure 4b) demonstrates that a basement thrust is a feasible solution for the structure of the High Zagros Fault in this region, but not a unique solution. Its throw appears to be comparable with blind thrusts inferred beneath the Simply Folded Belt to the south, i.e., in the kilometer range. The difference is that an exhumation threshold has been crossed at the higher elevations in the north. Once the Asmari Limestone is removed by erosion, it is rare for a mantle of Lower Tertiary and Upper Cretaceous clastics to survive on the anticline crests; erosion removes these softer rocks until the resistant Bangestan Group limestones are reached (Figure 5d).

[49] Whereas the High Zagros in the Bakhtyari Culmination (Figure 4a) contains abundant thrusts and exposes Palaeozoic strata and Hormuz Series salt, the equivalent zone in Fars consists of whaleback anticlines that are identical to counterparts in the Simply Folded Belt to the south (Figure 5d). Only the level of exhumation is different: deeper, but only on the scale of a kilometer or less. Large scale overthrusts and nappes of ophiolitic and/or suture zone rocks are not present until ≤30 km of the suture.

[50] The northeastern limit of major seismogenic thrusting lies within the Simply Folded Belt, up to 100 km south of the defined line of the High Zagros Fault. It does not correspond to any individual structure within the Simply Folded Belt, but there is a good correspondence with the 1250 m regional topographic contour [*Jackson and McKenzie*, 1984; *Talebian and Jackson*, 2004; *Nissen et al.*, 2011].

[51] Figure 4d is a restored section through the Fars line of Figure 4b. Northeast of Kuh-e Surmeh, the fold orientation is variable (Figure 2), which adds a degree of imprecision to the restored section. Shortening across the Simply Folded Belt is ~68 km or ~20%. The shortening value is higher than the Dezful Embayment/Bakhtyari Culmination line (55 km), although the percentage shortening is roughly the same. The Simply Folded Belt shortening estimate is in line with many previous estimates for the Zagros [see summaries in *Agard et al.*, 2011, *Vergés et al.*, 2011, and *Mouthereau et al.*, 2012], although much greater than those that assume little or no faulting in the cover [*Mouthereau et al.*, 2007; *Vergés et al.*, 2011]. As noted, the seismicity record requires faulting in the sedimentary cover with earthquakes of M~5 or above, and so we used this in our interpretation. Total shortening of the original Arabian plate is hard to determine because of greater uncertainty in the structure at depth, close to the suture zone.

[52] We tentatively estimate ~100 km shortening of basement across the entire Arabian plate, and ~120 km shortening of cover; the difference is due to the Radiolarite Group being stripped off its basement and transported farther southwest over the Arabian margin. These values are likewise in line with the restored section to the northwest but are also speculative because little is well constrained about the sub-surface structure near the suture zone. This estimate does not include any wholesale underthrusting of the thin, leading edge of the Arabian plate beneath Eurasia [*Paul et al.*, 2010].

[53] We have only interpreted basement faults in each cross-section where there is a distinct change in structural relief on either side of an exposed structure (e.g., Kuh-e Namak and Kuh-e Surmeh), consistent with seismicity data that suggest larger earthquakes in the Zagros occur in regions of greater structural relief, on faults that cut through the basement-cover boundary [*Nissen et al.*, 2011]. This can be seen as a minimum interpretation of basement involvement, but until more seismic data are available, it may not be possible to make more detailed interpretations. In any case, there are enough structural and seismicity data for basement involvement to call into question models that treat the Zagros as a critically tapered wedge above a basal decollement at the level of the Hormuz Series [e.g., *Ford*, 2004; *McQuarrie*, 2004]; it seems clear that as the basement is involved in the deformation, at very least, any critical taper model must include the basement [*Mouthereau et al.*, 2006]. There may be a compromise, whereby typical basement thrusts have moved only so far as to juxtapose the Hormuz Series rocks originally on either side of the Permian-Triassic rift. In the jargon of inversion tectonics, this is equivalent to saying that the null point of the inverted fault is at the level of the Hormuz Series.

4. Geomorphology

4.1. Topographic Profiles and Gradients

[54] Figures 6a and 6b are composite topographic profiles across the Dezful/Lursetan and Fars regions of the Zagros, following the line of the cross-sections in Figure 4, and showing the maximum, mean, and minimum elevations for a swath covering 50 km on either side of the central line. Elevations were obtained from version 4 of the Shuttle Radar Topography Mission (SRTM) dataset [*Jarvis et al.*

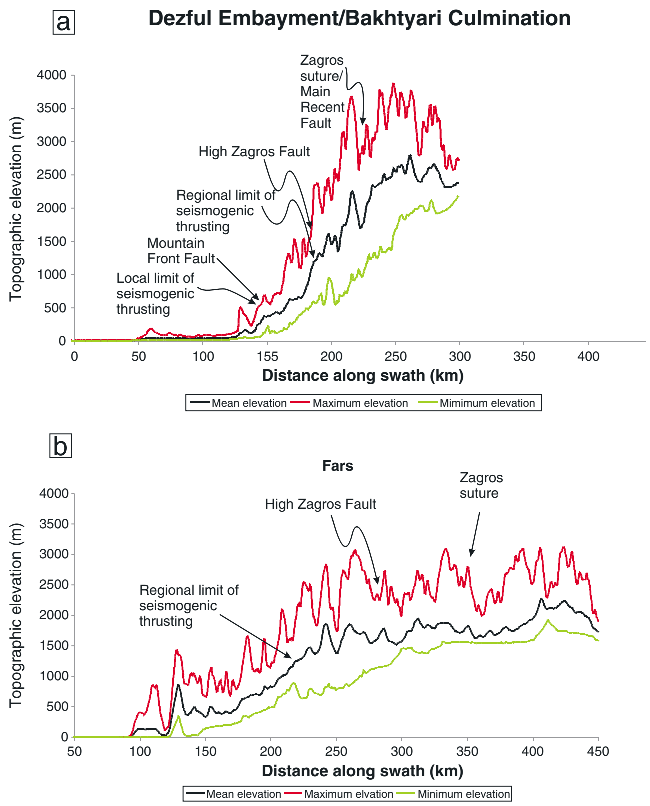

Figure 6. Topographic profiles across the Zagros for swaths extending 50 km to either side of the section lines of Figures 4a and 4b. Data are derived from the SRTM dataset.

2008]. SRTM data have a pixel size of 90 m, and specified vertical absolute accuracy of ≤16 m [*Rodriguez et al.* 2005]. Specific vertical relative accuracy is quoted as ≤10 m [*Rodriguez et al.* 2005].

[55] The interior of the Dezful Embayment is only a few 10s of meters above sea level. Average topographic gradient across the Dezful Embayment is 0.0004, i.e., a very low orogenic taper. Mean elevations were used for the calculation. Most of the elevation change (~2 km) in the Dezful/Bakhtyari Culmination section (Figure 6a) takes place northeast of the Mountain Front Fault, across ~100 km of the section, as far northeast as the line of the Main Recent Fault. The Main Recent Fault is commonly a prominent depression in topographic profiles across the Zagros, caused by the localized erosion along the steep fault trace, and the presence of pull-apart structures such as Borujerd [*Talebian and Jackson*, 2002]. The orogenic taper along the profile in Figure 6a between the Mountain Front Fault and the Main Recent Fault is 0.022. The final part of the section contains the highest elevations (typically ~2.5 km), and the drainage divide. Between the Main Recent Fault and the drainage divide, the taper is 0.012, and from the divide to the end of the section, it is 0.007, down to the northeast. Whereas the Mountain Front Fault is a clear change in the topography, corresponding to the start of the high gradient section noted above, the High Zagros Fault does not coincide with a significant topographic step. Nor is there a change in topography at the limit of seismogenic thrusting, either at the regional ~1250 m threshold or the local limit of <500 m along the topographic profile.

[56] The Fars profile (Figure 6b) shows a steady climb from the coast up to mean elevations of ~1750 m at ~180 km inland (250 km along the profile), with a taper of 0.010. The rest of the profile as far as the drainage divide at 425 km has a lower taper of 0.002; i.e., the mean elevations show a plateau topography. The difference between the maximum and minimum elevations decreases to the northeast of ~250 km along the profile, reflecting alluvial infilling of the synclinal valleys. Neither the High Zagros Fault nor the Main Zagros Reverse Fault corresponds to a pronounced topographic step; the change over from the higher to lower taper occurs southwest of the High Zagros Fault, and northeast of the 1250 m elevation contour that roughly marks the limit of seismogenic thrusting. The transition between the two parts of the section lies within the Simply Folded Belt and does not correspond to any major structure mapped at the surface. The swath includes externally drained regions as far as the High Zagros Fault, hence the steady climb in minimum elevation to this point.

[57] The maximum value profile in Figure 6b picks out the peaks of individual fold crests within both the Simply Folded Belt and the High Zagros. These are largely smoothed out in the mean value profile, because the swath width is greater than the length of any individual fold. A couple of folds are so long that their crests perturb the mean and minimum profiles.

[58] The rise in average elevations north of the seismogenic thrust limit (Figure 6b) occurs up to roughly the regional 1750 m elevation contour. It presumably involves an isostatic response to crustal thickening, but the precise mechanism for this thickening is literally cryptic: There is no signal from seismicity or available GPS coverage [*Hessami et al.*, 2006; *Walpersdorf et al.*, 2006; *Nissen et al.*, 2011] (Figure 1b). We return to this issue in section 5.

[59] To understand the topographic signatures in more detail, we have analyzed the variation in gradient and elevation across the Zagros, across the structural strike through both the Dezful Embayment and Fars regions. Data are derived from the SRTM v4 dataset [*Jarvis et al.* 2008].

[60] Figure 7 shows the gradients for the Dezful Embayment and adjacent areas of the Zagros, derived from the SRTM digital elevation data and processed using ArcGIS software. Low gradients across the Embayment (generally equivalent to slopes of <1°) contrast with higher gradients (>0.1, roughly equivalent to >10° slopes) within the Bakhtyari Culmination to its northeast, both within areas defined as the Simply Folded Belt and the High Zagros. This is a different pattern from that in either the Pusht-e Kuh arc to the northwest or the Fars region to the southeast. In both of these areas, a more consistent pattern of slopes applies from the deformation front to the northern limit of the High Zagros: High slopes over the spaced anticlines are separated by low slope regions of the synclines. The synclines are typically occupied by rivers, either draining externally or dammed. The valleys tend to be wider in the High Zagros than the Simply Folded Belt, both in the Pusht-e Kuh arc and in Fars (Figure 7). The strong carbonates exposed on the anticline flanks maintain high slopes, such that there is no discernible reduction in slope between the seismically active and inactive parts of the range (Figure 7).

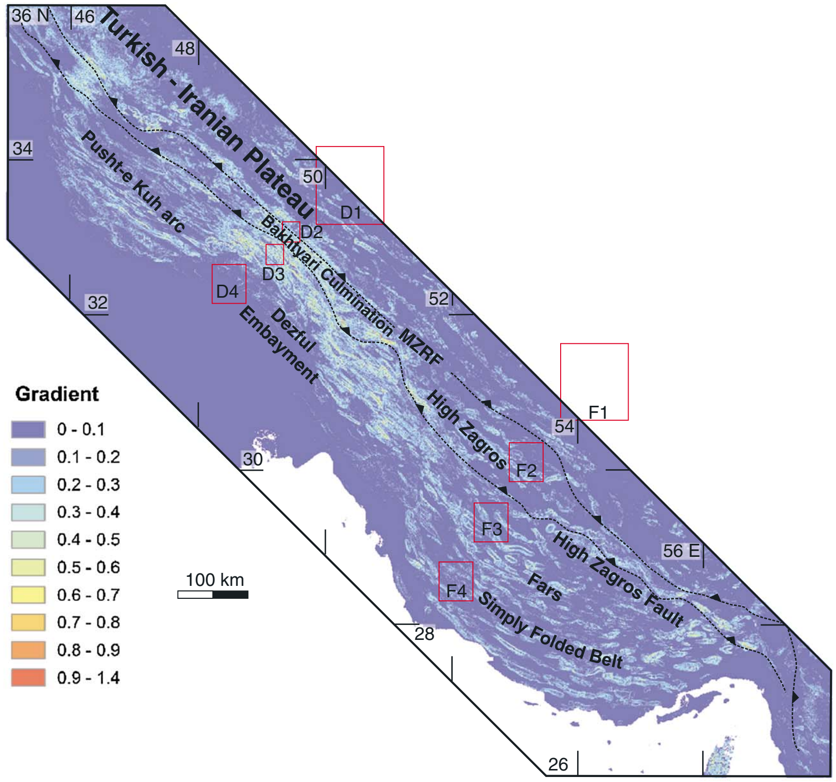

Figure 7. Gradient map for the Zagros, derived from the SRTM v4 dataset, using averages of 1 km windows. Highest gradients are predominately in the Bakhtyari Culmination.

[61] Figure 8 shows normalized frequency histograms for gradients within windows in the Turkish-Iranian Plateau and southwest across the suture in to the Zagros range. Data are shown in gradient divisions of 0.05. Four windows are used for two transects, across the Dezful Embayment/Bakhtyari Culmination and Fars; locations are shown on Figure 7. Window size is varied to make sure windows lie entirely within one or other of the zones of interest. Therefore, the windows northeast of the suture are $1° \times 1°$, while those to the southwest are either $0.5° \times 0.5°$, or $0.25° \times 0.25$ within the narrow Bakhtyari Culmination.

[62] Results for the Dezful/Bakhtyari Culmination windows, D1–D4, are shown in Figure 8a. There are low gradients within the Embayment itself (D4), as expected. Both the windows between the Mountain Front Fault and the Main Recent Fault (D2 and D3) show a roughly symmetrical distribution of values around a highest frequency gradient of 0.5–0.6. Not only are there high summits and steep slopes in these windows, but there are not significant plains along the valleys. Gradient distributions in region D3 (Simply Folded Belt above 1250 m) are intermediate between the distributions of D2 (High Zagros) and D4 (Dezful Embayment).

[63] Plateau morphology is only established northeast of the Main Recent Fault, where the frequency peak is at gradients of 0.05 (D1). We interpret these results to mean that crustal thickening has occurred northeast of the Mountain Front Fault, producing the steady increasing elevation. The mechanism is addressed in section 5.

[64] Results for the higher elevation Fars windows north of the limit of seismogenic thrusting (F1–F3) show peaks in frequency at gradients of 0.05 for each window (Figure 8b), with the frequency at this peak increasing from southwest to northeast. The lowest elevation window, within the seismogenic thrust sector of the Simply Folded Zone (F4), shows a frequency peak for gradients of 0.1, and a positively skewed distribution.

[65] The relief in the Fars region becomes more subdued beyond the limit of seismogenic thrusting. In this higher area, many of the synclinal basins are being infilled by alluvium, leading toward the construction of plateau geomorphology, including regions that are geologically part the Simply Folded Zone.

4.2. Drainage Patterns

[66] In this section, we look at the form of the main drainage basins that cover the Zagros, to see how they relate to the deformation patterns. *Mouthereau et al.* [2007] and *Ramsey et al.* [2008] have analyzed the interactions between rivers and individual folds in the Zagros. *Burberry et al.* [2008] presented a comprehensive map of drainage within the east of the Zagros.

[67] There are three main drainage basins in the Iranian Zagros (Figure 9) that drain through the rivers Karen, Mand, and Kul, and several smaller ones along the coast, such as Zohreh and Helleh [*Oberlander*, 1965]. Three internally drained areas occur within the Fars region: Niriz, Shiraz, and a collection of smaller basins we informally call Razak [*Mouthereau et al.*, 2007; *Walker et al.*, 2011], after the settlement within the area.

[68] Rivers in the Karun drainage basin (Figure 9) converge on the Dezful Embayment and join the Karun

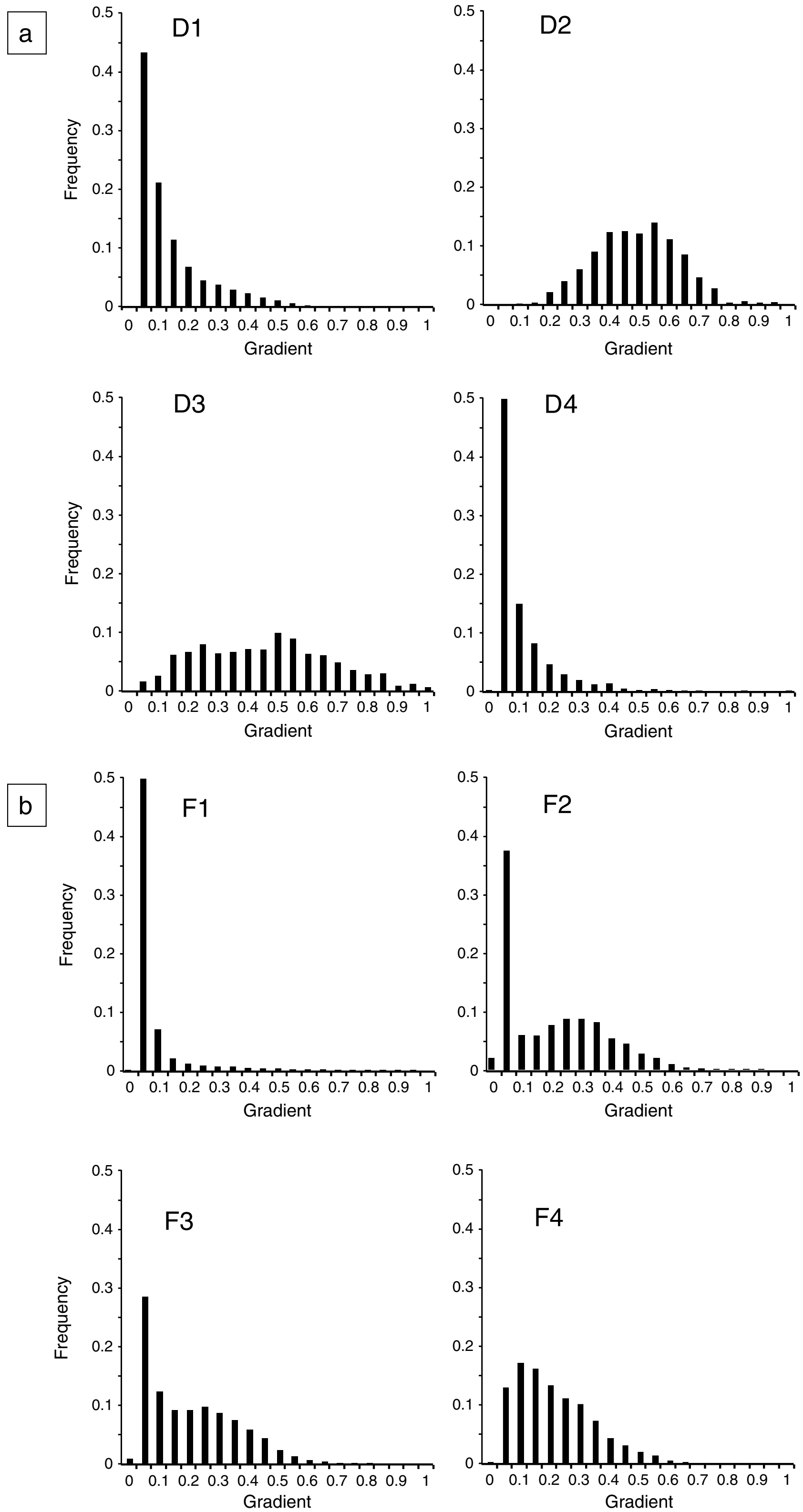

Figure 8. Normalized cumulative frequency plots for gradients of 1×1 km windows within the Dezful Embayment/Bakhtyari Culmination (a; regions D1–D4) and the Fars region (b; regions F1–F4). Data are derived from the SRTM v4 dataset.

River either within the Embayment or beyond its southwest boundary (Figure 9). The Karun River joins the Shatt Al-Arab ~80 km southeast of the confluence of the Tigris and the Euphrates. The latter two rivers are the axial drainage to the Zagros fold-and-thrust belt. The drainage pattern within the Dezful Embayment is consistent with its present

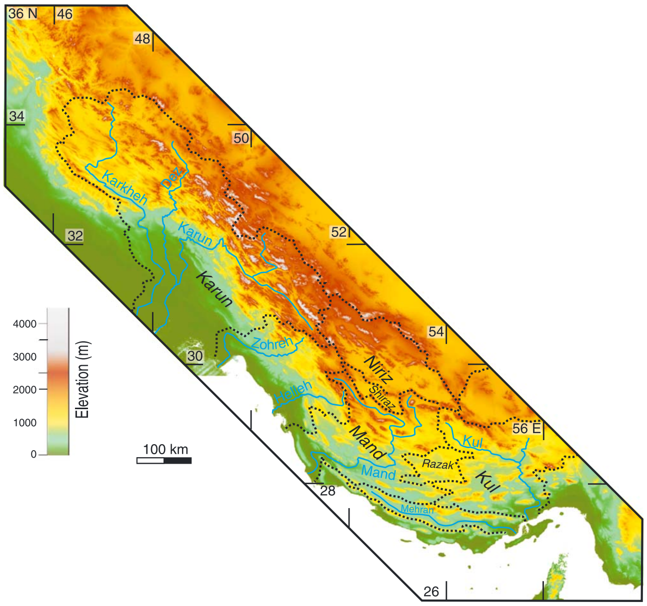

Figure 9. Drainage patterns of the Zagros. Note the internally drained areas both outside and within the zone of seismogenic thrusting.

low altitude, and its existence as a depocenter back to the Oligocene [*Koop and Stoneley*, 1982; *Motiei*, 1993].

[69] The northwest boundary to the Karun drainage basin is in the region of the Khanaqin Fault (Figure 2). The boundary between the Karun, Niriz, and Mand drainage basins lies close to the north-south Kareh Bas and Kazerun faults (Figure 2). Therefore, both of these prominent drainage divides occur at important structural boundaries within the Zagros. In both cases, the region of the drainage divide is marked by a change in relief, with higher ground and deeper exposure level to the east (Figure 9; *NIOC*, 1977).

[70] The Mand and Kul drainage basins are roughly equal in size and proportions (Figure 9). They are separated by the Razak region of internal drainage, roughly 10,000 km^2 in area [*Mouthereau et al*., 2007; *Walker et al*., 2011]. This region may have become isolated and through-going drainage defeated as the result of fold growth in the south [*Walker et al*., 2011]. *Walker et al*. [2011] and *Khadivi et al*. [2012] noted that sediment ponding would act as a process to raise average elevations in the lower elevation regions of internally drained basins and so encourage plateau development [*Sobel et al*., 2003]. Note that the Razak region lies below the regional 1250 m contour and within the area affected by seismogenic thrusts.

[71] There is less evidence of antecedent drainage in the Mand and Kul systems than in the Karun drainage basin, witnessed by the greater sinuosity of the Mand and Kul and their tributaries. There are more places where the rivers are diverted around the rising tips of folds [*Ramsey et al*., 2008]. This is possibly because discharge rates and sediment flux are lower in the Fars region than the Dezful Embayment/Bakhtyari Culmination [*Oberlander*, 1965], and topographic gradients are lower (section 4.1), such that the rivers lack sufficient stream power to cut through rising folds.

[72] The Iranian coastline of the Persian Gulf lacks major deltas southeast of the Tigris/Euphrates system and is relatively sediment starved, even at the mouths of the Mand and Kul Rivers (Figure 9). This in contrast to the Dezful Embayment, where progradation of the Karun has rapidly changed the shoreline of the Gulf [*Heyvaert and Baeteman*, 2007], possibly at rates of 20 m/a between 5000 and 2500 BP.

[73] Small drainage basins occupy the coastal strip of the eastern Zagros, enhancing the sediment starvation of this coast. Between them and the Mand and Kul drainage basins, there is an elongate drainage basin, occupied by the Mehran River, which flows west to east between two prominent anticlinal ridges for ~300 km [Figure 9; *Ramsey et al*., 2008]. The relative timings of drainage evolution are not certain, but it is likely that the Mehran drainage basin is a late stage feature, given its position near the frontal structures of the Zagros.

4.3. Landscape Evolution

[74] In section 3, we highlighted how the limit of seismogenic thrusting corresponds with the regional 1250 m elevation contour, but not a distinctive structural boundary at exposed levels. In section 4.1, we showed that the topographic profiles across the Zagros do not closely reflect this division between thrusting and non-thrusting parts of the range, with high regional topographic gradients beyond the limit of seismogenic thrusting. In this section, we describe local patterns of slope and incision in different parts of the Zagros.

[75] The development of plateau morphology in the Fars region is reflected in local patterns of deposition and incision (Figure 10a). Anticlines within the actively thrusting part of the Simply Folded Belt are associated with alluvial fans on their flanks (Figure 10b). These fans are traversed by braided river channels, which may be incised. Some anticlines are flanked by bajadas (Figure 10c). There is a lithological control on whether flanking fans are discrete or merged. Fold limbs that expose one or the other of the Asmari or Bangestan carbonates tend to be flanked by bajadas, formed from numerous small, parallel streams that flow down the fold limb. Kuh-e Mehmand is an example (Figure 10c). Where the Asmari Limestone has been breached but still forms the main topographic slope, drainage upstream of the Asmari Limestone outcrop coalesces into systems with enough stream power to penetrate the Asmari Limestone ridge that faces the fold axis (e.g., Kuh-e Namak, Figure 10b). Such rivers form wider-spaced drainage at lower elevations on the fold flanks.

[76] Rivers in the internally drained basins terminate in lakes such as Daryacheh-Ye Maharlu, 20 km southeast of Shiraz (Figure 10a). Significant lengths of the margins to these basins lack transverse alluvial fans. Sediment transported by axial rivers is aggrading, filling the basin topography. The anticlines are effectively being buried by the clastic sediments, even though the resistant carbonate lithologies maintain steep slopes (Figures 10d–10e).

[77] Our interpretation of this variation in landscape is that the alluvial fans on the frontal folds in the Zagros, which are actively deforming and undergoing surface uplift, are prone to tilting, incision, and cannibalisation by active river channels. Once an area has ceased seismogenic thrusting, it is smoothed out by erosion, leading first to the bajadas and in the final stage to the wide valleys and lakes within the High Zagros, without axial fans on the flanks of the subdued fold

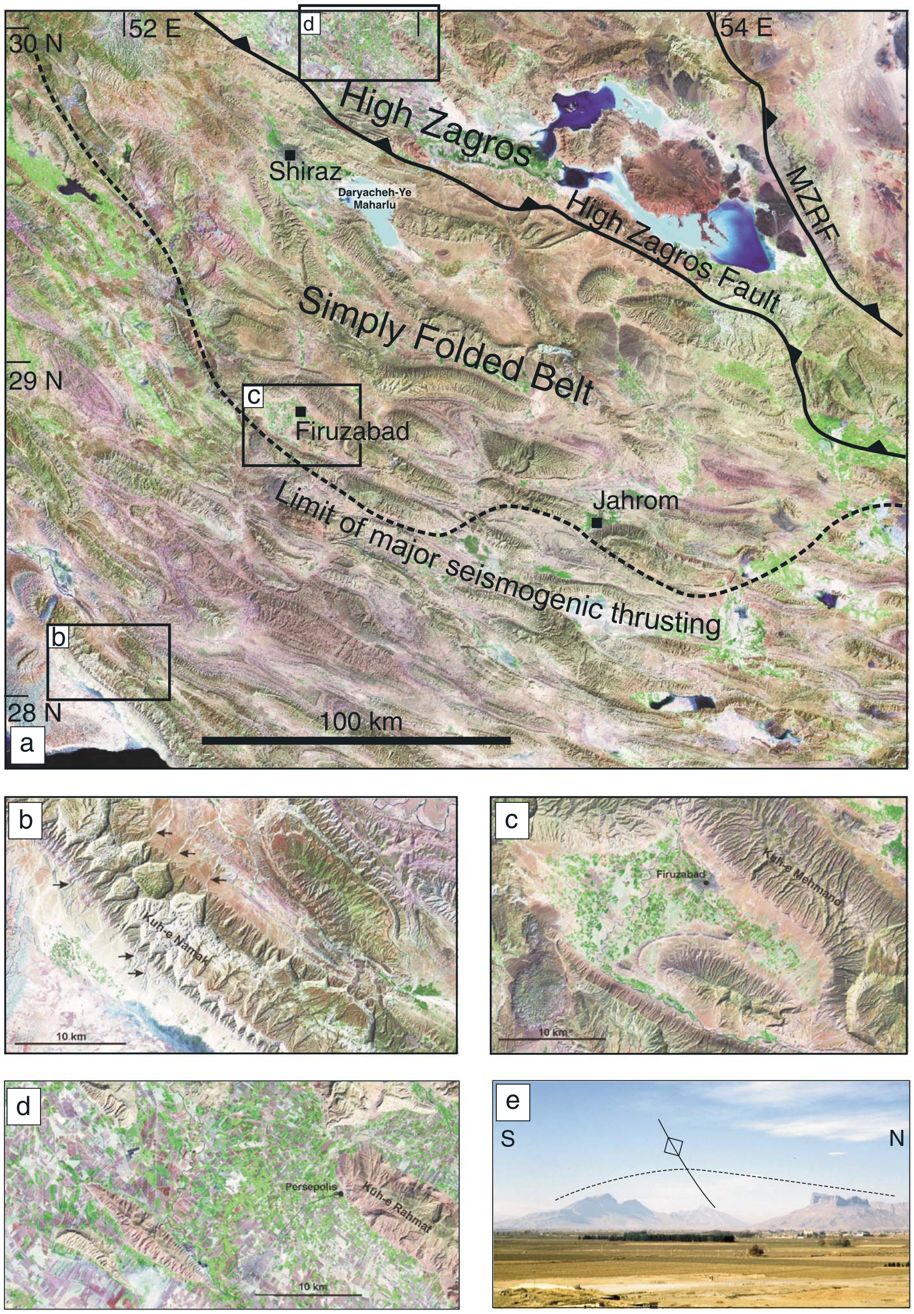

Figure 10. (a) Landsat (MrSID) mosaic of the Fars region (bands 7, 4, 2). (b) Close-up of Kuh-e Namak. Arrows highlight stream channels that have breached the carapace of Asmari Limestone. (c) Example of bajadas at the foot of ridges where the carbonate carapace is intact. (d) Drowned anticlines in the High Zagros: Valley floors are filling with alluvium in the internally drained Niriz basin. (e) View west of Persepolis (Takht-e Jamshid), showing alluvial plains beneath a dissected anticline of Bangestan Group limestones. The anticline has an open interlimb angle, typical of the strain in this part of the High Zagros, despite surface elevations of >1500 m.

topography. In this way, the lateral tectonic growth of the Turkish-Iranian Plateau toward the southwest is followed by its geomorphic growth. There is no fixed rate at which the time lag occurs, partly because there is a dependency on whether a region in the High Zagros keeps an efficient external drainage connection through to the rest of the range and the coast, or if it becomes internally drained [*Walker et al.*, 2011].

5. Discussion

[78] Several papers have addressed the topographic profiles across the Zagros in terms of implications for underlying detachment horizons, especially along the Hormuz Series at or near the basement/cover interface [*Ford*, 2004; *McQuarrie*, 2004; *Mouthereau et al.*, 2006; 2007]. The

width (~300 km) and elevation drop (~2 km) across the Fars sector of the Zagros combine (Figure 6b) to give it a low average gradient, previously interpreted in terms of an efficient Hormuz salt detachment allowing effective propagation of thrusting toward the foreland [*Ford*, 2004]. The problem with such interpretations lies in treating the Zagros as though there is a single gradient that applies across the range. This is not the case for either swath shown in Figure 6.

[79] An alternative explanation is as follows. Seismogenic thrusting takes place at elevations <1250 m, and associated crustal thickening produces the topographic gradient between sea level and this elevation [*Jackson and McKenzie*, 1984; *Nissen et al*., 2011]. Seismogenic thrusting ceases once the elevation gain is sufficient to generate enough gravitational potential energy to resist additional shortening [*Allen et al*., 2004]. GPS-derived velocity field data give support to this argument, as velocities drop off well southwest of the High Zagros (Figure 1b), although the current density of available data is not high enough to compare precisely to the seismicity and topography signals [*Hessami et al*., 2006; *Walpersdorf et al*., 2006]. But another process must be going on, as average elevations climb to ~1750 m in Fars, and ~2500 m northeast of the Dezful Embayment (Figure 6). If the extra elevation is a consequence of crustal shortening and thickening, via isostasy, part of this must happen by aseismic processes. However, low GPS-derived velocities, relative to central Iran, across the higher parts of the Simply Folded Belt and the High Zagros [*Hessami et al*., 2006; *Walpersdorf et al*., 2006] make this unlikely to be happening actively at upper crustal levels, in keeping with the reduction in thrust seismicity above the 1250 m regional elevation contour.

[80] We conclude that there is a contribution from aseismic and/or ductile crustal thickening of the basement beneath the higher parts of the Simply Folded Belt and the High Zagros, consistent with the increase in crustal thickness as the suture is approached from the southwest [*Paul et al*., 2006; 2010; *Rham*, 2009; *Mouthereau et al*., 2012].

[81] Underneath the suture zone and the Sanandaj-Sirjan Zone, high elevations may relate to underthrusting of the leading edge of the Arabian plate, with or without additional internal thickening of the crust [*Mouthereau et al*., 2007; *Paul et al*., 2010; *Agard et al*., 2011]. Southwest of the suture such underthrusting is not geometrically feasible because the region is within the Arabian plate. The same geometric argument applies to the idea that break-off of the subducted oceanic Neo-Tethyan slab could have caused surface uplift: This may apply north of the suture in regions such as eastern Turkey (given that any such break-off is likely to be a geologically recent phenomenon, *van Hunen and Allen*, 2011) but is unlikely to the south.

[82] Ductile shortening of the Arabian plate basement is a more plausible explanation of the elevations above the seismogenic limit of thrusting. This extra elevation above ~1250 m is effectively a topographic signal of depth-dependent strain during mountain building and plateau construction, and approximates to pure shear under the higher parts of the Zagros. It is analogous to models of depth-dependent continental extension, where different values for the extension factor, β, determined by upper crustal and whole crustal methods have been used to infer greater strain in the weak lower crust compared to the strong upper crust [e.g., *Driscoll and Karner*, 1998]. There is a further similarity to models of continental extension [e.g., *Jackson and White*, 1989], in that the Zagros need not be underlain by a single low-angle thrust, detaching everything above it from underlying basement. Instead, individual thrusts may tip out at the seismic-aseismic transition, passing into a region of diffuse, ductile deformation.

[83] We rule out a flexural effect from loading in the suture zone: The effective elastic thickness of the Arabian plate is ~15 km [*McKenzie and Fairhead*, 1997], which is much lower than the distance between the limit of seismogenic thrusting and the establishment of a low taper in the Fars region (~40 km, Figure 6b).

[84] Across much of the Turkish-Iranian Plateau farther north the elevated crust is neither especially thick nor actively shortening [~42 km across central Iran north of the Sanandaj-Sirjan Zone; *Paul et al*., 2006]. Support from mantle density variations may be a factor promoting the regional elevation in this region: *Maggi and Priestley* [2005] used surface wave tomography to show an area of low velocities in the upper mantle to depths of ~200 km below the Turkish-Iranian Plateau. The widespread, if patchy, Quaternary basaltic magmatism across the plateau implies partial melting, consistent with this idea [e.g., *Pearce et al*., 1990; *Kheirkhah et al*., 2009]. Such mantle may support elevated topography, above what is generated by the crustal thickness alone. However, *Allen et al*. [2011b] showed that some parts of the plateau near the Turkey/Iran border are characterized by relief that essentially pre-dates the Quaternary magmatism in the region, and slow rates of incision since the lavas flowed down the valleys (0.01–0.05 mm yr^{-1}).

[85] The cutoff in seismogenic thrusts in the Zagros occurs at lower elevations (~1250 m) than in the Himalayas and Tibet [~3.5 km; *England and Houseman*, 1988], and the maximum elevations of the Zagros and the Turkish-Iranian Plateau are also much lower (~2 km regional elevation compared with ~5 km for Tibet). It is not completely clear why this is the case. It may relate in some way to the strength of the crust in each area, with both the Arabian plate and Iran being weaker than India and Tibet [*England and Houseman*, 1988; *Maggi et al*., 2000]. There is a striking similarity between both plateaux, notwithstanding the differences in absolute values of elevation: In both regions, there is a regional gain in elevation above the seismogenic limit of thrusting, implying that another process is involved. In Tibet, this extra elevation has been explained by loss of the lower lithosphere and consequent isostatic uplift [*England and Houseman*, 1988]. However, recent work on the deep structure of Tibet indicates that a thick lithosphere is in place under much if not all of the plateau [*Priestley and McKenzie*, 2006; *Chen et al*., 2010], with obvious implications for delamination models. Nor can underthrusting of the Indian plate be responsible for the excess elevation at all the plateau margins: It is not simply a phenomenon confined to a linear front across Tibet. We therefore speculate that depth-dependent crustal shortening is responsible for the extra elevation of the Tibetan plateau, above the 3500 m limit of seismogenic thrusting.

[86] New crustal-scale cross-sections help define the regional structure of the Zagros. In the eastern Zagros (Fars), the whole fold-and-thrust belt is ~300 km across strike and has a similar structural style throughout, albeit with a deeper erosion level in the High Zagros than the Simply Folded

Belt. In the northwest Zagros (Lurestan/Khuzestan provinces), the fold-and-thrust belt is narrower and includes a low-deformation zone within the southwest of the Simply Folded Belt: the Dezful Embayment. This region has resisted late Cenozoic deformation better than its surroundings and acts as a major depocenter. Deformation is correspondingly more intense northeast of the Dezful Embayment in the Bakhtyari Culmination, presumably to maintain roughly constant strain across the Zagros as a whole. This is reflected in the geomorphology of this area, with the highest elevations and steepest slopes in the whole of the Zagros (Figures 7 and 8). As a consequence of the Dezful Embayment, lateral plateau growth is more limited at these longitudes than to the east in Fars, where plateau growth has migrated furthest south in the Zagros.

[87] Several factors in Fars create low exhumation rates, which lead to more rapid crustal thickening for any given shortening, and therefore crust that more rapidly reaches the elevation threshold at which seismogenic thrusting stops. These factors are as follows: a semi-arid climate and consequent low erosion rates; a regional salt detachment and/or inherited basement faults—both of which encourage deformation to propagate toward the foreland; and resistant carbonate lithologies at both Tertiary and Cretaceous levels, which impede erosion and exhumation.

[88] These ideas are summarized in Figure 11, which shows seismogenic thrusting occurring in a belt across the Zagros that at any one time is confined to the region below ~1250 m elevation. As crustal thickness builds up, the location of this belt migrates toward the foreland, over time including areas previously unaffected by significant thrusting.

[89] There is no abrupt change in the landscape of the Zagros to match the cutoff in active, seismogenic thrusting. But there is a reduction in topographic gradient across the hinterland parts of the range, and a tendency to form internally drained basins. Sediment filling these basins enhances the smooth topography in these regions, as anticlines are gradually buried in alluvium [*Walker et al*., 2011]. Even in areas with external drainage, the semi-arid climate keeps the rivers relatively small, while limestone gorges choke off sediment flux and pond it within synclinal basins. Such inefficient erosion and sediment transport reduces the average gradient of the Zagros, even though summits of carbonate-cored anticlines resist erosion and maintain high relief above the surrounding plains.

[90] The way in which seismogenic thrusting relates to topography in the Zagros, rather than major structural boundaries, is more in keeping with models of tectonics that highlight its diffuse nature [continuum approach, e.g., *Liu and Bird*, 2008], rather than microplate models [e.g., *Reilinger et al*., 2006]. Microplate models may honor the kinematic data but do not capture the incremental changes in the distribution of strain over time, nor the possibility that lower and upper crustal deformation are decoupled. By its very nature, aseismic, probably ductile, basement strain is difficult to detect. Topographic signals, such as the elevation climb of the Zagros detailed in this paper, may be one of the best insights in areas of active deformation.

[91] The structure of the Zagros is also very relevant to critical taper models of fold-and-thrust belt growth [*Dahlen*, 1990] but raises some problems: Steep thrusts that are present and active across wide areas of the range and involve

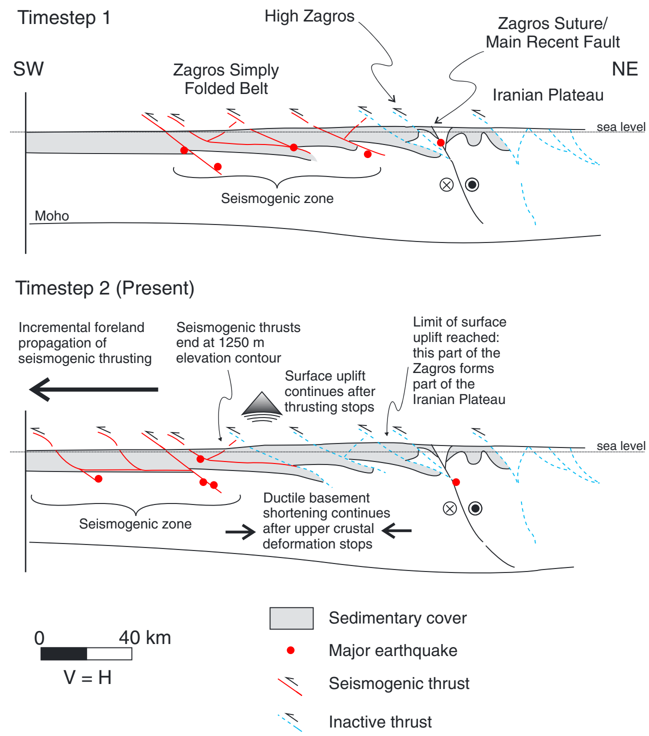

Figure 11. Schematic model for the evolution of thrusting in the Zagros. At any one time, several thrusts are seismically active within the range (shown by schematic timestep 1), but this seismic belt moves toward the south, keeping to the foreland of the regional 1250 m elevation contour (timestep 2: active tectonics). It is speculated that aseismic, ductile basement shortening continues northeast of the seismogenic faults but that such deformation also reaches a limit once sufficient crustal thickness and elevation are reached, and the area effectively forms part of the Turkish-Iranian Plateau.

basement do not easily fit the idea of a single, low-angle detachment at the base of a deforming zone. Nor is there a single topographic slope that applies across any given cross-section; rising topography beyond the limit of seismogenic thrusting implies a decoupling of upper crustal (brittle) and lower crustal (ductile) deformation, but one that is ultimately limited by the buoyancy force associated with elevated crust, rather than a requirement to maintain a critical angle for an entire orogenic wedge.

[92] **Acknowledgments.** We thank the Geological Survey of Iran for their support. Part of this work was started when MBA and EJB were at CASP, Cambridge. This work was supported by the Natural Environment Research Council [grant number NE/H021620/1]. We thank Olivier Lacombe, an anonymous reviewer, and the editors for their helpful comments on the first version of the paper.

References

Adams, A., R. Brazier, A. Nyblade, A. Rodgers, and A. Al-Amri (2009), Source parameters for moderate earthquakes in the Zagros mountains with implications for the depth extent of seismicity, *Bull. Seis. Soc. Am.*, *99*, 2044–2049.

Agard, P., J. Omrani, L. Jolivet, and F. Mouthereau (2005), Convergence history across Zagros (Iran): Constraints from collisional and earlier deformation, *Int. J. Earth Sci.*, *94*, 401–419.

Agard, P., J. Omrani, L. Jolivet, H. Whitechurch, B. Vrielynck, W. Spakman, P. Monie, B. Meyer, and R. Wortel (2011), Zagros orogeny: A subduction-dominated process, *Geol. Mag.*, *148*, 692–725.

Ahmadhadi, F., O. Lacombe, and J. M. Daniel (2007), Early reactivation of basement faults in Central Zagros (SW Iran): Evidence from pre-folding fracture populations in Asmari formation and lower tertiary paleogeography, in Thrust Belts and Foreland Basins: From Fold Kinematics to Hydrocarbon Systems, edited by O. Lacombe, J. Lave, F. Roure and J. Vergés, pp. 205–228, Springer-Verlag, New York, NY.

Alavi, M. (1994), Tectonics of the Zagros orogenic belt of Iran: New data and interpretations, *Tectonophysics*, *239*, 211–238.

Alavi, M. (2007), Structures of the Zagros fold-thrust belt in Iran, *Am. J. Sci.*, *307*, 1064–1095.

Allen, M. B., and H. A. Armstrong (2008), Arabia-Eurasia collision and the forcing of mid Cenozoic global cooling, *Palaeogeogr. Palaeoclim. Palaeoecol.*, *265*, 52–58.

Allen, M. B., and M. Talebian (2011), Structural variation along the Zagros and the nature of the Dezful Embayment, *Geol. Mag.*, *148*, 911–924.

Allen, M., J. Jackson, and R. Walker (2004), Late Cenozoic reorganization of the Arabia-Eurasia collision and the comparison of short-term and long-term deformation rates, *Tectonics*, *23*, TC2008, doi: 2010.1029/2003TC001530.

Allen, M. B., M. Kheirkhah, M. H. Emami, and S. J. Jones (2011a), Right-lateral shear across Iran and kinematic change in the Arabia-Eurasia collision zone, *Geophys. J. Int.*, *184*, 555–574.

Allen, M. B., D. F. Mark, M. Kheirkhah, D. Barfod, M. H. Emami, and C. Saville (2011b), $^{40}Ar/^{39}Ar$ dating of Quaternary lavas in northwest Iran: Constraints on the landscape evolution and incision rates of the Turkish-Iranian Plateau, *Geophys. J. Int.*, *185*, 1175–1188.

Ameen, M. S. (1991), Possible forced folding in the Taurus-Zagros belt of northern Iraq, *Geol. Mag.*, *128*, 561–584.

Aqrawi, A. A. M., J. C. Goff, A. D. Horbury, and F. N. Sadooni (2010), The Petroleum Geology of Iraq, Scientific Press Ltd, Beaconsfield, U.K.

Authemayou, C., D. Chardon, O. Bellier, Z. Malekzadeh, E. Shabanian, and M. R. Abbassi (2006), Late Cenozoic partitioning of oblique plate convergence in the Zagros fold-and-thrust belt (Iran), *Tectonics*, *25*, TC3002, doi:10.1029/2005tc001860.

Authemayou, C., O. Bellier, D. Chardon, L. Benedetti, Z. Malekzade, C. Claude, B. Angeletti, E. Shabanian, M. R. Abbassi (2009), Quaternary slip-rates of the Kazerun and the Main Recent Faults: Active strike-slip partitioning in the Zagros fold-and-thrust belt, *Geophys. J. Int.*, *178*, 524–540.

Axen, G. J., P. S. Lam, M. Grove, D. F. Stockli and J. Hassanzadeh (2001), Exhumation of the west-central Alborz Mountains, Iran, Caspian subsidence, and collision-related tectonics, *Geology*, *29*, 559–562.

Babaie, H. A., A. Babaei, A. M. Ghazi, and M. Arvin (2006), Geochemical, Ar-40/Ar-39 age, and isotopic data for crustal rocks of the Neyriz ophiolite, Iran, *Can. J. Earth Sci.*, *43*, 57–70.

Bahroudi, A., and H. A. Koyi (2003), Effect of spatial distribution of Hormuz salt on deformation style in the Zagros fold and thrust belt: An analogue modelling approach, *J. Geol. Soc. Lond.*, *160*, 719–733.

Baker, C., J. Jackson, and K. Priestley (1993), Earthquakes on the Kazerun Line in the Zagros Mountains of Iran: Strike-slip faulting within a fold-and-thrust belt, *Geophys. J. Int.*, *115*, 41–61.

Ballato, P., C. E. Uba, A. Landgraf, M. R. Strecker, M. Sudo, D. F. Stockli, A. Friedrich, and S. H. Tabatabaei (2011), Arabia-Eurasia continental collision: Insights from late Tertiary foreland-basin evolution in the Alborz Mountains, northern Iran, *Geol. Soc. Am. Bull.*, *123*, 106–131, doi:10.1130/B30091.1.

Barnes, J. B., and T. A. Ehlers (2009), End member models for Andean Plateau uplift, *Earth-Sci. Rev*, *97*, 105–132.

Berberian, M. (1995), Master “blind” thrust faults hidden under the Zagros folds: Active basement tectonics and surface morphotectonics, *Tectonophysics*, *241*, 193–224.

Beydoun, Z. R., M. W. Hughes Clarke, and R. Stoneley (1992), Petroleum in the Zagros Basin: A late Tertiary foreland basin overprinted onto the outer edge of a vast hydrocarbon-rich Paleozoic-Mesozoic passive-margin shelf, in Foreland Basins and Foldbelts, edited by R. MacQueen and D. Leckie, pp. 309-339, AAPG Memoir 55, Tulsa, Oklahoma, USA.

Blanc, E. J.-P., M. B. Allen, S. Inger, and H. Hassani (2003), Structural styles in the Zagros Simple Folded Zone, Iran, *J. Geol. Soc. Lond.*, *160*, 400–412.

Bosold, A., W. Schwarzhans, A. Julapour, A. R. Ashrafzadeh, and S. M. Ehsani (2005), The structural geology of the High Central Zagros revisited (Iran), *Pet. Geosci.*, *11*, 225–238.

Braud, J. (1970), Les formations du Zagros dans la région de Kermanshah (Iran) et leur rapport structuraux, *Comptes Rendus de l’Académie des Sciences 271*, 1241–1244.

Burberry, C. M., J. W. Cosgrove, and J. G. Liu (2008), Spatial arrangement of fold types in the Zagros Simply Folded Belt, Iran, indicated by landform morphology and drainage pattern characteristics, *J. Maps*, *2008*, 417–430.

Carruba, S., C. R. Perotti, R. Buonaguro, R. Calabro, R. Carpi, and M. Naini (2006), Structural pattern of the Zagros fold-and-thrust belt in the Dezful Embayment (SW Iran), in Styles of Continental Contraction, edited by S. Mazzoli and R. W. H. Butler, pp. 11–32, Geological Society of America Special Paper 414, Boulder, Colorado, USA.

Casciello, E., J. Vergés, E. Saura, G. Casini, N. Fernandez, E. Blanc, S. Homke, and D. W. Hunt (2009), Fold patterns and multilayer rheology of the Lurestan Province, Zagros Simply Folded Belt (Iran), *J. Geol. Soc. Lond.*, *166*, 947–959.

Chen, W. P., M. Martin, T. L. Tseng, R. L. Nowack, S. H. Hung, and B. S. Huang (2010), Shear-wave birefringence and current configuration of converging lithosphere under Tibet, *Earth Planet. Sci. Lett.*, *295*, 297–304.

Clark, M. K., K. A. Farley, D. W. Zheng, Z. C. Wang, and A. R. Duvall (2010), Early Cenozoic faulting of the northern Tibetan Plateau margin from apatite (U-Th)/He ages, *Earth Planet. Sci. Lett.*, *296*, 78–88.

Dahlen, F. A (1990), Critical taper model of fold-and-thrust belts and accretionary wedges, *Ann. Rev. Earth Planet. Sci.*, *18*, 55–99.

Dalmayrac, B., and P. Molnar (1981), Parallel thrust and normal faulting in Peru and constraints on the state of stress, *Earth Planet. Sci. Lett.*, *55*, 473–481.

DeMets, C., R. G. Gordon, D. F. Argus, and S. Stein (1994), Effects of recent revisions to the geomagnetic time scale on estimates of current plate motions, *Geophys. Res. Lett.*, *21*, 2191–2194.

Dhont, D., J. Chorowicz and P. Luxey (2006), Anatolian escape tectonics driven by Eocene crustal thickening and Neogene-Quaternary extensional collapse in the eastern Mediterranean region, in Postcollisional Tectonics in the Mediterranean Region and Asia, edited by Y. Dilek and S. Pavlides, pp. 441–462, Geological Society of America Special Paper 409, Boulder, Colorado, USA.

Driscoll, N. W., and G. D. Karner (1998), Lower crustal extension across the Northern Carnarvon basin, Australia: Evidence for an eastward dipping detachment, *J. Geophys. Res.*, *103*, 4975–4991.

Edgell, H. S. (1991), Proterozoic salt basins of the Persian Gulf area and their role in hydrocarbon generation, *Precambrian Res.*, *54*, 1–14.

England, P. C., and D. McKenzie (1982), A thin viscous sheet model for continental deformation, *Geophys. J. Royal Astron. Soc.*, *70*, 295–321.

England, P. C., and G. A. Houseman (1988), The mechanics of the Tibetan plateau, *Phil. Trans. R. Soc. Lond.*, *326*, 301–319.

Fakhari, M. D., G. J. Axen, B. K. Horton, J. Hassanzadeh, and A. Amini (2008), Revised age of proximal deposits in the Zagros foreland basin and implications for Cenozoic evolution of the High Zagros, *Tectonophysics*, *451*, 170–185.

Farzipour-Saein, A., A. Yassaghi, S. Sherkati, and H. Koyi (2009), Mechanical stratigraphy and folding style of the Lurestan region in the Zagros Fold-Thrust Belt, Iran, *J. Geol. Soc. Lond.*, *166*, 1101–1115.

Ford, M. (2004), Depositional wedge tops: Interaction between low basal friction external orogenic wedges and flexural foreland basins, *Basin Res.*, *16*, 361–375.

Frizon de Lamotte, D., C. Raulin, N. Mouchot, J. C. Wrobel-Daveau, C. Blanpied, and J. C. Ringenbach (2011), The southernmost margin of the Tethys realm during the Mesozoic and Cenozoic: Initial geometry and timing of the inversion processes, *Tectonics*, *30*, Tc3002, doi:10.1029/2010tc002691.

Gansser, A. (1992), The enigma of the Persian salt dome intrusions, *Eclogae Geologicae Helveticae*, *85*, 825–846.

Garzione, C. N., P. Molnar, J. Libarkin, and B. MacFadden (2006), Rapid late Miocene rise of the Bolivian Altiplano: Evidence for removal of mantle lithosphere, *Earth Planet. Sci. Lett.*, *241*, 543–556.

Gavillot, Y., G. J. Axen, D. F. Stockli, B. K. Horton and M. K. Fakhari (2010), Timing of thrust activity in the High Zagros fold-thrust belt, Iran, from (U-Th)/He thermochronometry, *Tectonics*, *29*, TC4025, doi: 10.1029/2009TC002484.

Ghazi, A. M., and A. A. Hassanipak (1999), Geochemistry of subalkaline and alkaline extrusives from the Kermanshah ophiolite, Zagros Suture Zone, Western Iran: Implications for Tethyan plate tectonics, *J. Asian Earth Sci.*, *17*, 319–332.

Gok, R., H. Mahdi, H. Al-Shukri, and A. J. Rodgers (2008), Crustal structure of Iraq from receiver functions and surface wave dispersion: Implications for understanding the deformation history of the Arabian-Eurasian collision, *Geophys. J. Int.*, *172*, 1179–1187.

Guest, B., D. F. Stockli, M. Grove, G. J. Axen, P. S. Lam, and J. Hassanzadeh (2006), Thermal histories from the central Alborz Mountains, northern Iran: Implications for the spatial and temporal distribution of deformation in northern Iran, *Geol. Soc. Am. Bull.*, *118*, 1507–1521.

Hatzfeld, D., and P. Molnar (2010), Comparisons of the kinematics and deep structures of the Zagros and Himalaya and of the Iranian and Tibetan plateaus and geodynamic implications, *Rev. Geophys*, *48*, Rg2005, doi:10.1029/2009rg000304.

Hessami, K., F. Nilforoushan, and C. J. Talbot (2006), Active deformation within the Zagros Mountains deduced from GPS measurements, *J. Geol. Soc. Lond.*, *163*, 143–148.

Hessami, K., H. A. Koyi, C. J. Talbot, H. Tabasi, and E. Shabanian (2001), Progressive unconformities within an evolving foreland fold-thrust belt, Zagros Mountains, *J. Geol. Soc. Lond.*, *158*, 969–981.

Heyvaert, V. M. A., and C. Baeteman (2007), Holocene sedimentary evolution and palaeocoastlines of the Lower Khuzestan plain (southwest Iran), *Marine Geol.*, *242*, 83–108.

Holt, P. J., M. B. Allen, J. van Hunen, and H. M. Bjørnseth (2010), Lithospheric cooling and thickening as a basin forming mechanism, *Tectonophysics*, *495*, 184–194.

Homke, S., J. Vergés, M. Garcés, H. Emami, and R. Karpuz (2004), Magnetostratigraphy of Miocene-Pliocene Zagros foreland deposits in the front of the Push-e Kush Arc (Lurestan Province, Iran), *Earth Planet. Sci. Lett.*, *225*, 397–410.

Iran Oil Operating Companies (1969), Geological Map of South-West Iran, scale 1:1,000,000, Iran Oil Operating Companies, Tehran, Iran.

Jackson, J. (2001), Living with earthquakes: Know your faults, *J. Earthquake Eng.*, *5*, 5–123.

Jackson, J., and D. McKenzie (1984), Active tectonics of the Alpine-Himalayan belt between western Turkey and Pakistan, *Geophys. J. Royal Astron. Soc.*, *77*, 185–264.

Jackson, J. and D. McKenzie, (1988), The relationship between plate motion and seismic moment tensors, and the rates of active deformation in the Mediterranean and Middle East, *Geophys. J. Royal Astron. Soc.*, *93*, 45–73.

Jackson, J. A. and N. J. White, (1989), Normal faulting in the upper crust—Observations from regions of active continental extension. *J. Struct. Geol.*, *11*, 15–36.

James, G. A., and J. G. Wynd (1965), Stratigraphic nomenclature of the Iranian oil consortium agreement area, *Bull. Am. Assoc. Petrol. Geol.*, *49*, 2182–2245.

Jarvis, A., H. I. Reuter, A. Nelson, and E. Guevara (2008), Hole-filled SRTM for the globe version 4, available from the CGIAR-CSI SRTM 90m Database (http://srtm.csi.cgiar.org).

Kaviani, A., A. Paul, E. Bourova, D. Hatzfeld, H. Pedersen and M. Mokhtari (2007), A strong seismic velocity contrast in the shallow mantle across the collision zone (Iran), *Geophys. J. Int.*, *171*, 399–410.

Kent, P. E. (1979), The emergent Hormuz salt plugs of southern Iran, *J. Pet. Geol.*, *2*, 117–144.

Keskin, M. (2003), Magma generation by slab steepening and breakoff beneath a subduction-accretion complex: An alternative model for collision-related volcanism in Eastern Anatolia, Turkey, *Geophys. Res. Lett.*, *30*, art. No. 8046, doi:10.1029/2003GL018019.

Khadivi, S., F. Mouthereau, J. C. Larrasoana, J. Vergés, O. Lacombe, E. Khademi, E. Beamud, M. Melinte-Dobrinescu, and J. P. Suc (2010), Magnetochronology of synorogenic Miocene foreland sediments in the Fars arc of the Zagros Folded Belt (SW Iran), *Basin Res.*, *22*, 918–932.

Khadivi S., F. Mouthereau, J. Barbarand, T. Adatte and O. Lacombe (2012), Constraints on paleodrainage evolution induced by uplift and exhumation on the southern flank of the Zagros-Iranian Plateau, *J. Geol. Soc. London*, *169*, 83–97.

Kheirkhah, M., M. B. Allen, and M. Emami (2009), Quaternary syn-collision magmatism from the Iran/Turkey borderlands, *J. Volc., Geotherm. Res.*, *182*, 1–12.

Koop, W. J., and R. Stoneley (1982), Subsidence history of the Middle East Zagros Basin, Permian to Recent, *Phil. Trans. R. Soc. Lond.*, *A305*, 149–168.

Lacombe O., F. Mouthereau, S. Kargar, and B. Meyer (2006), Late Cenozoic and modern stress fields in the western Fars (Iran): Implications for the tectonic and kinematic evolution of Central Zagros, *Tectonics*, *25*, TC1003, doi:10.1029/2005TC001831.

Liu, Z., and P. Bird (2008), Kinematic modelling of neotectonics in the Persia-Tibet-Burma orogen, *Geophys. J. Int.*, *172*, 779–797.

Liu-Zeng, J., P. Tapponnier, Y. Gaudemer, and L. Ding (2008), Quantifying landscape differences across the Tibetan plateau: Implications for topographic relief evolution, *J. Geophys. Res.*, *113*, F04018, doi: 10.1029/2007jf000897.

Llewellyn, P. (1972), *Ahwaz*, scale 1:250,000, Iranian Oil Operating Companies, Tehran, Iran.

Llewellyn, P. (1973), *Dezful*, scale 1:250,000, Iranian Oil Operating Companies, Tehran, Iran.

Maggi, A., and K. Priestley (2005), Surface waveform tomography of the Turkish-Iranian plateau, *Geophys. J. Int.*, *160*, 1068–1080.

Maggi, A., J. A. Jackson, K. Priestley, and C. Baker (2000), A re-assessment of focal depth distributions in southern Iran, the Tien Shan and northern India: Do earthquakes really occur in the continental mantle? *Geophys. J. Int.*, *143*, 629–661.

Mazhari, S. A., F. Bea, S. Amini, J. Ghalamghash, J. F. Molina, P. Montero, J. H. Scarrow, and I. S. Williams (2009), The Eocene bimodal Piranshahr massif of the Sanandaj-Sirjan Zone, NW Iran: A marker of the end of the collision in the Zagros orogen, *J. Geol. Soc. Lond.*, *166*, 53–69.

McClusky, S., et al. (2000), Global Positioning System constraints on plate kinematics and dynamics in the eastern Mediterranean and Caucasus, *J. Geophys. Res.*, *105*, 5695–5719.

McKenzie, D., and D. Fairhead (1997), Estimates of the effective elastic thickness of the continental lithosphere from Bouguer and free air gravity anomalies, *J. Geophys. Res.*, *102*, 27523–27552.

McQuarrie, N. (2004), Crustal scale geometry of the Zagros fold-thrust belt, Iran, *J. Struct. Geol.*, *26*, 519–535.

McQuillan, H. (1991), The role of basement tectonics in the control of sedimentary facies, structural patterns and salt plug emplacements in the Zagros foldbelt of southwest Iran, *J. Southeast Asian Earth Sci.*, *5*, 453–463.

Meyer, B., F. Mouthereau, O. Lacombe, and P. Agard (2006), Evidence of Quaternary activity along the Deshir Fault: Implication for the Tertiary tectonics of central Iran, *Geophys. J. Int.*, *164*, 192–201.

Molinaro, M., H. Zeyen, and X. Laurencin (2005), Lithospheric structure beneath the south-eastern Zagros Mountains, Iran: Recent slab break-off? *Terra Nova*, *17*, 1–6.

Morley, C. K., B. Kongwung, A. A. Julapour, M. Abdolghafourian, M. Hajian, D. Waples, J. Warren, H. Otterdoom, K. Srisuriyon, and H. Kazemi (2009), Structural development of a major late Cenozoic basin and transpressional belt in central Iran: The Central Basin in the Qom-Saveh area, *Geosphere*, *5*, 325–362.

Motiei, H. (1993), Stratigraphy of Zagros, 536 pp., Geological Survey of Iran, Tehran.

Mouthereau F. (2011), Timing of uplift in the Zagros belt/Iranian plateau and accommodation of late Cenozoic Arabia-Eurasia convergence, *Geol. Mag.*, *148*, 726–738.

Mouthereau, F., O. Lacombe, and B. Meyer (2006), The Zagros folded belt (Fars, Iran): Constraints from topography and critical wedge modelling, *Geophys. J. Int.*, *165*, 336–356.

Mouthereau, F., J. Tensi, N. Bellahsen, O. Lacombe, T. De Boisgrollier, and S. Kargar (2007), Tertiary sequence of deformation in a thin-skinned/thick-skinned collision belt: The Zagros folded belt (Fars, Iran), *Tectonics*, *26*, Tc5006, doi:10.1029/2007tc002098.

Mouthereau, F., O. Lacombe and J. Vergés (2012), Building the Zagros collisional orogeny: Timing, strain distribution and the dynamics of Arabia/Eurasia convergence, *Tectonophysics*, *532*, 27–60, doi:10.1016/j.tecto2012.01.022.

Murris, R. J. (1980), Middle East: Stratigraphic evolution and oil habitat, *Bull. Am. Assoc. Petrol. Geol.*, *64*, 597–618.

National Iranian Oil Company (1975), *Geological Map of Iran Sheet 4 South-West Iran*, scale 1:1,000,000, National Iranian Oil Company, Tehran.

National Iranian Oil Company (1977), *Geological Map of Iran Sheet 5 South-Central Iran*, scale 1:1,000,000, National Iranian Oil Company, Tehran.

Nemati, M., and A. Yassaghi (2010), Structural characteristics of the transitional zone from internal to external parts of the Zagros orogen, Iran, *J. Asian Earth Sci.*, *39*, 161–172.

Nissen, E., M. Ghorashi, J. Jackson, B. Parsons and M. Talebian (2007), The 2005 Qeshm Island earthquake (Iran)—A link between buried reverse faulting and surface folding in the Zagros Simply Folded Belt? *Geophys. J. Int.*, *171*, 326–338.

Nissen, E., M. Tatar, J. A. Jackson, and M. B. Allen (2011), New views on earthquake faulting in the Zagros fold-and-thrust belt of Iran, *Geophys. J. Int.*, *186*, 928–944.

Nissen, E., F. Yamini-Fard, M. Tatar, A. Gholamzadeh, E. Bergman, J. R. Elliott, J. A. Jackson, and B. Parsons (2010), The vertical separation of mainshock rupture and microseismicity at Qeshm island in the Zagros fold-and-thrust belt, Iran, *Earth Planet. Sci. Lett.*, *296*, 181–194.

O' Brien, C. A. E. (1957), Salt diapirism in South Persia, *Geologie en Mijnbouw*, *19*, 357–376.

Oberlander, T. (1965), The Zagros streams, 168 pp. Syracuse Geographical Series 1.

Okay, A. I., M. Zattin and W. Cavazza (2010), Apatite fission-track data for the Miocene Arabia-Eurasia collision, *Geology*, *38*, 35–38.

Oveisi, B., J. Lave, P. van der Beek, J. Carcaillet, L. Benedetti, and C. Aubourg (2009), Thick- and thin-skinned deformation rates in the central Zagros simple folded zone (Iran) indicated by displacement of geomorphic surfaces, *Geophys. J. Int.*, *176*, 627–654.

Paul, A., A. Kaviani, D. Hatzfeld, J. Vergne, and M. Mokhtari (2006), Seismological evidence for crustal-scale thrusting in the Zagros mountain belt (Iran), *Geophys. J. Int.*, *166*, 227–237.

Paul, A., D. Hatzfeld, A. Kaviani, M. Tatar, and C. Péquegnat (2010), Seismic imaging of the lithospheric structure of the Zagros mountain belt (Iran), in Tectonic and Stratigraphic Evolution of Zagros and Makran during the Mesozoic-Cenozoic, edited by P. Leturmy and C. Robin, pp. 5–18, Geol. Soc., Lond., Special Publications, 330, London, UK.

Pearce, J. A., J. F. Bender, S. E. Delong, W. S. F. Kidd, P. J. Low, Y. Guner, F. Sargolu, Y. Yilmaz, S. Moorbath, and J. G. Mitchell (1990), Genesis of collision volcanism in eastern Anatolia, Turkey, *J. Volc., Geotherm. Res.*, *44*, 189–229.

Pirouz, M., G. Simpson, A. Bahroudi, and A. Azhdari (2011), Neogene sediments and modern depositional environments of the Zagros foreland basin system, *Geol. Mag.*, *148*, 838–853.

Priestley, K., and D. McKenzie (2006), The thermal structure of the lithosphere from shear wave velocities, *Earth Planet. Sci. Lett.*, *244*, 285–301.

Ramsey, L. A., R. T. Walker, and J. Jackson (2008), Fold evolution and drainage development in the Zagros mountains of Fars province, SE Iran, *Basin Res.*, *20*, 23–48.

Rezaeian, M., A. Carter, N. Hovius, and M. B. Allen (2012), Cenozoic exhumation of the Alborz Mountains, Iran: New constraints from low-temperature chronometry. *Tectonics*, *31*, TC2004, doi: 10.1029/2011TC002974.

Reilinger, R., et al. (2006), GPS constraints on continental deformation in the Africa-Arabia-Eurasia continental collision zone and implications for the dynamics of plate interactions, *J. Geophys. Res.*, *111*, B05411, doi 05410.01029/02005JB004051.

Rham, D. J. (2009), The crustal structure of the middle East, PhD thesis, Univ. of Cambridge, Cambridge, U.K.

Rodriguez, E., C. S. Morris, J. E. Belz, E. C. Chapin, J. M. Martin, W. Daffer, and S. Hensley (2005), An assessment of the SRTM topographic products, *Jet Propulsion Laboratory*, *Pasadena*, *California*, *Technical Report D-31639*, 1–143.

Roustaei, M., E. Nissen, M. Abbassi, A. Gholamzadeh, M. Ghorashi, M. Tatar, F. Yamini-Fard, E. Bergman, J. Jackson, and B. Parsons (2010), The 2006 March 25 Fin earthquakes (Iran)—Insights into the vertical extents of faulting in the Zagros Simply Folded Belt, *Geophys. J. Int.*, *181*, 1275–1291.

Rowley, D. B., and B. S. Currie (2006), Palaeo-altimetry of the late Eocene to Miocene Lunpola basin, central Tibet, *Nature*, *439*, 677–681.

Sella, G. F., T. H. Dixon, and A. Mao (2002), REVEL: A model for recent plate velocities from space geodesy, *J. Geophys. Res.*, *107*, art. no. 2081 doi 2010.1029/2000JB000033.

Şengör, A. M. C., and W. S. F. Kidd (1979), Post-collisional tectonics of the Turkish-Iranian plateau and a comparison with Tibet, *Tectonophysics*, *55*, 361–376.

Şengör, A. M. C., D. Altiner, A. Cin, T. Ustaomer, and K. J. Hsu (1988), Origin and assembly of the Tethyside orogenic collage at the expense of Gondwana Land, in *Gondwana and Tethys*, edited by M. G. Audley-Charles and A. Hallam, pp. 119–181, *Geol. Soc., Lond.*, Special Publications, 37, London, UK.

Sepehr, M., and J. W. Cosgrove (2004), Structural framework of the Zagros fold-thrust belt, Iran, *Mar. Pet. Geol.*, *21*, 829–843.

Sepehr, M., J. Cosgrove, and M. Moieni (2006), The impact of cover rock rheology on the style of folding in the Zagros fold-thrust belt, *Tectonophysics*, *427*, 265–281.

Setudehnia, A. (1978), The Mesozoic sequence in south-west Iran and adjacent areas, *J. Pet. Geol.*, *1*, 3–42.

Sherkati, S., and J. Letouzey (2004), Variation of structural style and basin evolution in the central Zagros (Izeh zone and Dezful Embayment), Iran, *Mar. Pet. Geol.*, *21*, 535–554.

Sherkati, S., J. Letouzey and D. Frizon de Lamotte (2006), Central Zagros fold-thrust belt (Iran): New insights from seismic data, field observation, and sandbox modelling, *Tectonics 25*, TC4007. doi:10.1029/2004TC001766.

Sobel, E. R., G. E. Hilley, and M. R. Strecker (2003), Formation of internally drained contractional basins by aridity-limited bedrock incision, *J. Geophys. Res.*, *108*, art. no. 2344 doi 2310.1029/2002JB001883.

Soleimany, B., and F. Sabat (2010), Style and age of deformation in the NW Persian Gulf, *Pet. Geosci.*, *16*, 31–39.

Stöcklin, J. (1968), Structural history and tectonics of Iran: A review, *Bull. Am. Assoc. Petrol. Geol.*, *52*, 1129–1258.

Szabo, F., and A. Kheradpir (1978), Permian and Triassic stratigraphy, Zagros Basin, south-west Iran, *J. Pet. Geol.*, *1*, 57–82.

Talbot, C. J., and M. Alavi (1996), The past of a future syntaxis across the Zagros, in Salt Tectonics, edited by G. I. Alsop, D. J. Blundell and I. Davison, pp. 89-109, Geol. Soc., Lond., Special Publications, 100, London, UK.

Talebian, M., and J. Jackson (2002), Offset on the Main Recent Fault of NW Iran and implications for the late Cenozoic tectonics of the Arabia-Eurasia collision zone, *Geophys. J. Int.*, *150*, 422–439.

Talebian, M., and J. Jackson (2004), A reappraisal of earthquake focal mechanisms and active shortening in the Zagros mountains of Iran, *Geophys. J. Int.*, *156*, 506–526.

Tapponnier, P., Z. Q. Xu, F. Roger, B. Meyer, N. Arnaud, G. Wittlinger, and J. S. Yang (2001), Oblique stepwise rise and growth of the Tibet plateau, *Science*, *294*, 1671–1677.

Tatar, M., D. Hatzfeld, and M. Ghafory-Ashtiany (2004), Tectonics of the Central Zagros (Iran) deduced from microearthquake seismicity, *Geophys. J. Int.*, *156*, 255–266.

Van Hunen, J. and M. B. Allen (2011), Continental collision and slab break-off: A comparison of 3-D numerical models with observations, *Earth Planet. Sci. Lett.*, *302*, 27–37

Vergés, J., E. Saura, E. Casciello, M. Fernandez, A. Villasenor, I. Jimenez-Munt, and D. Garcia-Castellanos (2011), Crustal-scale cross-sections across the NW Zagros belt: Implications for the Arabian margin reconstruction, *Geol. Mag.*, *148*, 739–761.

Vernant, P., et al. (2004), Contemporary crustal deformation and plate kinematics in Middle East constrained by GPS measurements in Iran and northern Iran, *Geophys. J. Int.*, *157*, 381–398.

Vincent, S. J., A. C. Morton, A. Carter, S. J. Gibbs, and T. C. Barabadze, (2007), Oligocene uplift of the Western Greater Caucasus: An effect of initial Arabia-Eurasia collision, *Terra Nova*, *19*, 160–166.

Walker, R., J. Jackson, and C. Baker (2004), Active faulting and seismicity of the Dasht-e-Bayaz region, eastern Iran, *Geophys. J. Int.*, *157*, 265–282.

Walker, R. T., L. A. Ramsey, and J. Jackson (2011), Geomorphic evidence for ancestral drainage patterns in the Zagros Simple Folded Zone and growth of the Iranian plateau, *Geol. Mag.*, *148*, 901–910.

Walpersdorf, A., D. Hatzfeld, H. Nankali, F. Tavakoli, F. Nilforoushan, M. Tatar, P. Vernant, J. Chery, and F. Masson (2006), Difference in the GPS deformation pattern of north and central Zagros (Iran), *Geophys. J. Int.*, *167*, 1077–1088.

Yamato, P., B. J. P. Kaus, F. Mouthereau, and S. Castelltort (2011), Dynamic constraints on the crustal-scale rheology of the Zagros fold belt, Iran, *Geology*, *39*, 815–818.